

**The Holocene:
an
Environmental
History**

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NEIL ROBERTS

THE HOLOCENE

AN ENVIRONMENTAL HISTORY

THIRD EDITION



WILEY Blackwell

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*To Glyn and Betty for putting me on the right road, and to Sylvie and
Sara for keeping me there.*

THE HOLOCENE AN ENVIRONMENTAL HISTORY

THIRD EDITION

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Technical boxes

Dating the onset of the Holocene

Tephrochronology

Dividing up Late Quaternary time

Lymnology

Subdividing the Holocene

Morphological changes in plant domesticates

Ölzi – the ice man

Ruddiman's early anthropogenic greenhouse hypothesis

The Anthropocene

Climate from tree rings

Preface to the third edition

Since 1986, when the first edition of this book was published, Holocene research has grown enormously, above all linked to wider concerns about global environmental change. Developments over recent decades have included new techniques of field and laboratory analysis, as well as remarkable discoveries such as Ötzi, the ice man. Important new data syntheses have been published and stimulating conceptual ideas have emerged. Some of these have prompted lively debate, such as Bill Ruddiman's early anthropogenic greenhouse impact. In a related vein, Paul Crutzen and Gene Stoermer proposed that we have now entered a new geological epoch dominated by global human impact – the Anthropocene. If this comes to be accepted, then the Holocene will already have ended! These and other developments demanded to be incorporated in a new edition of the book in order to bring it up to date. More than 500 new references have been added, so that the Bibliography now includes three times the number of items in the first edition – such is the pace at which Holocene research has been produced and published in the last three decades. At the same time, I have sought to retain the best of the older classic literature by pioneering scholars such as Ed Deevey, Winifred Pennington, Sir Harry Godwin and Johannes Iversen – ‘giants on whose shoulders we stand’.

Important though these changes are, the biggest shift since the publication of this book's first edition is the realisation that we are now living through an era of human-induced climate change. This is no longer the possibility it was in the mid-1980s, but a reality. Every single year since this book was first published in 1986 has been warmer than the twentieth-century average, and during that time the mean global temperature has risen by around 0.5°C. The Holocene provides a vital long-term context for current climate warming, not least because it tells us that the planet is moving into previously uncharted territory. Realisation of the reality of human-induced climate change has produced a scientific response, notably since the establishment of the IPCC in 1988, but so far much less in the way of political consensus, let alone action. With hindsight, the major change that has been occurring in the Earth System in recent decades is now clear to see, even if it was not evident at the time. Where it will take us in the decades to come is much harder to foretell.

Note on chronology

As in the second edition of this book, calendar years before present (Cal. yr BP) are used throughout, sometimes abbreviated to ka BP (thousand years before

present).

Glossary items appear in **bold blue** in the text

Acknowledgements

Holocene research is inherently an interdisciplinary venture, drawing its source material from disciplines as diverse as biology, archaeology, geomorphology/geology and climatology. All of these are included within the realm of this book, but the resulting synthesis aims to 'look to the past to interpret the present', and as such is above all a work of geography. Geography seems particularly well fitted to a synthesising role, with its concern for variations in nature and culture across space and through time. As well as the debt to my home discipline, I am enormously grateful to colleagues who have over the years influenced ideas which appear in this book or who have commented on the text, among them Phil Barker, Rick Battarbee, Keith Bennett, the late Sytze Bottema, Jane Bunting, Robin Butlin, Chris Caseldine, Dan Charman, the late Denis Cosgrove, Pete Downs, Dan Engstrom, Ralph Fyfe, Françoise Gasse, Roland Gehrels, Andrew Goudie, Dick Grove, David Harris, Gordon Hillman, John Kutzbach, Henry Lamb, Scott Mooney, Rewi Newnham, Frank Oldfield, Bas Payne, Reinhard Pienitz, Jim Ritchie, Arlene Rosen, Bill Ruddiman, Heikki Seppä, the late Andrew Sherratt, Alayne Street-Perrott, Matt Telfer, Charles Turner, Claudio Vita-Finzi, Tom Webb, Herb Wright and Giovanni Zanchetta. Errors and omissions are of course my own. Thanks are due to the Director and researchers at the CNRS Laboratoire de Géologie du Quaternaire (now CEREGE) in Aix-Marseille, where the writing of the first edition of the book took place when I was on sabbatical leave. I am similarly grateful to Ian Hodder, the Stanford Archaeology Center and to a Blaustein fellowship for support during a subsequent sabbatical visit at Stanford University where much of the third edition was drafted. Thanks go to Val Pheby and Gwynneth Barnwell for secretarial help; Ann Tarver, Peter Robinson, Erica Millwain, Linda Dawes and Tim Absalom for cartography; the late Vernon Poulter for photographic assistance; and the Departments of Geography at Loughborough and Plymouth Universities. Staff at Wiley Blackwell Publishers, past and present, have also been instrumental in ensuring that this and previous editions finally appeared, and my thanks go to Jill Landeryou, Emma Gotch, Kelvin Matthews, Leander Shrimpton, Delia Sandford, Ian Francis and John Davey. My family showed great tolerance while I have been engaged in writing, and re-writing, this tome, and to them I shall be ever grateful. Finally, a special debt is owed to 'la lumière et les paysages du Muy' which supplied much of the original inspiration for this book.

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Introduction

Are the moors and downs of the British Isles natural landscapes or were they created by human agency? To what extent has soil loss through erosion increased on cultivated land compared with areas of natural woodland? And is the recorded rise in global temperatures since 1980 a result of the current increase in greenhouse gases in the atmosphere, or is it just part of climate's natural tendency to vary through time? Answers to problems such as these are often sought by monitoring contemporary environmental processes. For example, Gordon Wolman (1967) measured sediment yields from forested and cultivated land in the eastern United States and found that erosion was up to eight times higher for the latter. Alternatively, data from gauged field stations may be used, although the short time period of observation often proves to be a handicap. Most meteorological stations, for example, only have records going back to the first half of the twentieth century (Jones et al., 1999), and this makes it hard to identify any long-term trends of warming or cooling.

In fact, neither monitoring nor gauged records are likely on their own to provide complete solutions to the problems posed. This is because a longer-term view is required. An average human lifespan is so much shorter than the millennia of natural history that we tend to be aware only of short-term variations in the environment – the wet summer, the late spring, the 'record' flood. What we are much less aware of are slower, subtler changes such as alterations in the floristic composition of woodland, the silting up of estuaries and the advance and retreat of glaciers. Yet only a long-term perspective could tell us that Britain's downs and moors were transformed from their original woodland before written history even began (Simmons, 2003).

Sources of information on past environments

For historic times, documentary sources can sometimes provide reliable

observations on the former state of the natural environment. Tax records have been used to indicate late seventeenth- and early eighteenth-century climatic deterioration in southern Norway (Grove, 2004), while maps and legal documents show the changing position of Spurn Point spit on the east coast of England during the last 300 years (De Boer, 1964). Among the best historical sources are the early United States federal land surveys which mapped the pre-existing forest and grassland vegetation, and even recorded their species composition, before laying out land boundaries. Documentary data of this sort form an important part of historical ecology which applied to past flora and fauna (Sheail, 1980), historical geology when related to changes in the physical landscape such as rivers (Hooke and Kain, 1982) and historical climatology when linked to former climates (Bradley and Jones, 1992; Brázdil et al., 2005). In practice, the distinction between these sub-disciplines is often an arbitrary one.

Written records are, however, restricted to literate cultures, and as late as AD 1500, they existed only in Europe, Asia and North Africa. There are consequently long periods of human history for which recorded observations are absent, termed prehistory by archaeologists. Moreover, written history covers very different time spans in different parts of the world. Whereas written history began around 5000 years ago in Mesopotamia with the Sumerians, and 2000 years ago in Britain with Julius Caesar, it only started in the 1930s in the highlands of Papua New Guinea when aircraft brought the first European contact. Prehistory has come to be associated in Western thought with all that is remote, both in time and in cultural affinity, to twenty-first-century life. On the other hand, to a New Guinea highlander (or a Maori or black Zimbabwean), prehistory represents a direct cultural heritage which ended only a few generations ago. All of this means that the attitudes towards the natural world of many past societies have either gone unrecorded, or have appeared in written form only through the eyes of others. Ethnobotanical and other studies, however, have revealed a remarkable indigenous knowledge of the natural environment and its uses by modern non-literate hunter-gatherers, peasant farmers and nomads (Cunningham, 2001).

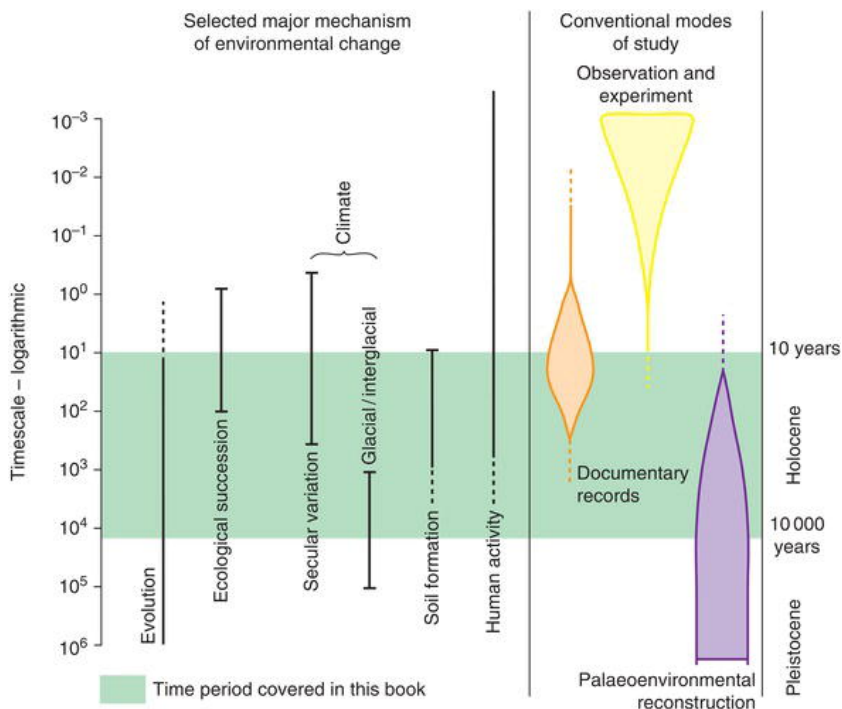
It is not only from ancient times and distant places that historical data on the natural environment prove to be deficient. Old men may recall how, when in their youth, fine catches of salmon were taken from Scandinavian rivers now devoid of fish, but because no one recorded the pH of the water until acid rain became a problem during the 1970s, it is difficult to know whether the salmon were eliminated by [acidification](#) or simply overfished. Availability of documentary sources therefore tends to inhibit consideration of non-literate regions such as Papua New Guinea and problems such as acidification, which do not appear in historical accounts. The example of megafaunal extinctions (see Chapter 3) serves to illustrate the point. When told that much African wildlife is threatened by extinction, our reaction is one of horror. Yet extinctions of even greater magnitude have occurred within the time span of human history considered in this book, notably in the Americas and Australia where up to 90% of all large mammals were lost. However, this ecological crisis remains known only to a narrow audience of specialists because no contemporary written accounts of it were handed down to posterity.

This book aims to recount and try to explain changes in the natural world through time, including those in human–environment relations. If a truly cross-

cultural perspective is to be taken on this, then we need to escape from the restrictions imposed by information derived solely from contemporary or documented historical sources. This is not to deny the critical importance of such sources, but while they have been employed extensively in studies of environmental relations (e.g. Glacken, 1967; Worster, 1993; McNeill, 2000), other sources have not. Without a time machine in which to return to the past, we have instead to rely on proxy evidence, including that from palaeo-science, which provides environmental histories, and archaeology, which provides human histories. These two sources sometimes come together as **environmental archaeology**, in which plant remains, animal bones and sediments from archaeological sites are studied to reconstruct their past economy and environment (Branch et al., 2005). These subjects share in common many techniques such as **radiocarbon dating** and pollen analysis, and they are often investigated together to form interdisciplinary research projects.

For the moment, it is sufficient to note some of the strengths and weaknesses of the palaeoenvironmental approach (Rymer, 1980; Oldfield, 2005). One of its weaknesses is that it cannot reconstruct attitudes to the natural world by former human societies in the way that historical sources can. Medieval cosmologies which placed 'man' in a holistic relationship with the natural world, for example, would be scarcely comprehensible without texts to explain them (Cosgrove, 2008). Environmental archaeology, with its concern for site economies and food remains, inclines the investigator towards an economic view of past human life (e.g. Higgs, 1972). This bias towards economic and away from social and cultural explanations is most easily countered where the archaeological past meets the anthropological present, for instance, with the Hopi Indians of the American southwest (Butzer, 1982). In the case of prehistoric Europe, it is not so easy. Even so, we occasionally get glimpses of past symbolic and ritual activities linked to the natural world; for example, pollen of meadowsweet (*Filipendula ulmaria*) from Bronze Age burials in Scotland demonstrates how floral tributes were placed next to their dead (Tipping, 1994; Clarke, 1999).

Figure 1.1 Mechanisms and modes of studying environmental change over different timescales (modified from Oldfield, 1983). Reproduced with permission of the Geographical Association.



On the other hand, one of the great advantages of archaeology and palaeo-science is their ability to identify long-term patterns of cultural or environmental development (Caseldine and Turney, 2010; Caseldine, 2012). This is especially important because the timescales of adjustment in many environmental systems span centuries or millennia rather than individual years, via processes such as soil maturation and ecological succession (see Figure 1.1). In the case of palaeo-science, environmental reconstruction can be applied to more recent as well as to long-term changes. Most lake sediments, for instance, have continued to accumulate up to the present day, and they can therefore be used to give a historical dimension to such contemporary environmental problems as pollution of freshwaters. The continuous nature of most palaeoecological records provides what Frank Oldfield (1977) aptly described as ‘a true continuum of insight’. This enables us to establish whether environmental changes have been episodic or gradual and to identify baseline conditions if they existed. These techniques, unlike written records, have the further advantage of being as applicable in New Guinea as they are in New England.

Nature and society

All of the world’s landscapes and ecosystems are products of the natural and

cultural processes that have shaped them over time to bring them to their present state. But what are those processes? Over the last one million years, a major factor has been climatic change, with the climate oscillating between Pleistocene glacials and interglacials. Only after the end of the last glaciation, around 12 000 years ago, did the world's climates and environments take on a recognisably modern form. Secular climatic variations of smaller magnitude and shorter duration have continued up to the present day. But whereas most environmental changes brought about by natural agencies have diminished in amplitude as one moves forward in time, there is one set of processes which has done precisely the reverse; human impact on the environment has increased progressively through time as *Homo sapiens* has been transformed from hunter-gatherer to city-dweller.

The oldest and simplest human mode of production is hunting, fishing and gathering (h-f-g), described by Mandel (1969, p. 34) as a 'primitive economy...in which the results are so meager that they must be shared to avoid death by starvation'. In fact, a wide range of anthropological studies suggest that far from eking out a meagre existence in a hostile environment, h-f-g groups probably represent the original affluent society (Sahlins, 1974; Simmons, 2008). Old World h-f-g populations are represented archaeologically by **Palaeolithic** and **Mesolithic** cultures.

Hunting and gathering was followed in the process of cultural evolution by peasant farming, the two being distinguished by the latter's exploitation of domestic plants and animals. The adoption of agriculture by h-f-g groups, termed the **Neolithic** revolution by Vere Gordon Childe, was a decisive moment in human history. Undeniably, it changed the basis of human relations with nature. The basic form of production in h-f-g and subsistence agricultural economies is simple and communal, with no systematic expropriation of surplus labour. This is not true, however, of more complex agricultural economies which supported feudal, classical and 'hydraulic' societies. The resulting states and civilizations typically – although not always – brought with them literacy and consequently the start of written history. Finally, the last two centuries of human history have witnessed the rise of industrial, and arguably post-industrial, economies and the global expansion of Western European culture. Human impact on the Earth system in recent centuries has become so large that it has been proposed that it should be designated as a new geological era – the Anthropocene (see **Technical Box IX**). Each stage in this social evolution of humankind has seen an increase in control over our relationship with nature (Simmons, 1996, 2008). As we have evolved from hunting to agriculture to industry and beyond, so human impact upon the environment has apparently come to counterbalance or even replace environmental influence over human affairs.

A great strength of an evolutionary approach to human–environment relations is that it is historically mediated. However, consideration of the historical dynamic is usually focused on the human half of the partnership (e.g. Diamond, 1997). The forces of nature are all too often viewed as an essentially passive backcloth against which human history is acted out. Although this environmental backcloth may change from one scene to another, it plays little or no active part in its own right. Perhaps the sky will be permitted to vary to show storm, snow or sun as a token gesture to the fact that nature is not static, but the forests and streams will be ever present, the natural landscape constant. It is above all this

ahistorical view of nature that this book seeks to challenge. Climate, forests and rivers have their histories too.

By employing a long-term perspective, it becomes possible for us to ask interesting and important research questions, such as the following:

- When, where and how has our planet moved from a being nature- to human-dominated?
- Can the past provide us with meaningful targets for restoring damaged environments?
- What lessons can the Holocene offer us about societal responses to climatic adversity and other environmental crises?
- Has the Earth system crossed any ‘tipping points’ since the time of the Last Ice Age, and if so, what determined these non-linear responses?
- Can the Holocene tell us anything about the possible future course of climate, and what might happen to ice, ocean and biota on planet Earth?

This book not only asks but also attempts to answer these and other questions in the chapters that follow, and particularly in the concluding chapter.

The significance of the Holocene

The period over which most of these cultural and environmental changes have taken place is the [Holocene](#). The Holocene – or post-glacial – epoch provides the time frame for this book, which is organised in chronological fashion, starting with the last glacial-to-interglacial transition and then working towards the present day.

During the late Pleistocene, 25 000–12 000 years ago, the glacial climate made the Earth cold and unfamiliar. Canada lay buried beneath several kilometres of ice, Europe was largely devoid of forests, and the southeast Asian islands were joined to form a single land mass – these are some of the changes described in Chapter 3. This chapter also discusses human evolution and the enigmatic – and not unrelated – mass extinction of megafauna in Australia and the Americas. The huge shift in climate which occurred at the end of the last glaciation was important not only in its own right but also because it indirectly controlled many other parts of the Earth system. Post-glacial adjustments of plant and animal distributions, sea levels, geologic and soil-forming processes are discussed in Chapter 4, along with the changes in climate that took place during the first half of the Holocene. This time period, between 11 700 and 6000 years ago, is when we might look to if we were to search for nature’s primeval, virginal baseline state – one free from ‘significant’ human disturbance. However, if it ever existed, this condition was to end with the [domestication](#) of plants and animals. The emergence of farming and its initial impact upon forest ecosystems are considered in Chapter 5. Chapter 6 focuses on later Holocene environments,

notably those produced by complex agroecosystems such as those which developed around the Mediterranean basin and Mesoamerica. It also includes an account of the landscape history of Britain and Ireland. Chapter 7 moves forward to the last millennium; it discusses the impact of human populations upon land use and ecology as we expanded our occupation of the planet's lands and oceans during historical times and the role of industrial capitalism upon pollution of the atmosphere and freshwaters. Included too is the historical background to recent human-induced changes in atmospheric greenhouse gas concentrations, with its potential for altering the global climate. Finally, Chapter 8 brings these threads together to offer an overview of our changing relationship with nature, with implications for how we might put into practice our responsibility as environmental stewards.

But before embarking on a natural history of the Holocene, we need to establish how this remarkable story can be told. For this reason, Chapter 2 of this book is devoted to describing the main techniques that are available for us to reconstruct and date past environmental changes.

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Reconstructing Holocene environments

Dating the past

Dating techniques are fundamental to an understanding of the natural and cultural changes which have taken place during the Holocene. Without them, events such as the **Neolithic** agricultural revolution would float aimlessly in time. Perhaps worse, without independent dating methods, we would be forced to depend on environmental and archaeological evidence to provide our chronologies, thus robbing this evidence of much of its potential meaning (Vita-Finzi, 1973).

Curiously, the most obvious reason for wanting to date the past may be the least important, that is, the desire to know how old something is simply for its own sake. It may stagger the imagination to think that there are trees living today in California's White Mountains which were saplings before Stonehenge was completed, but the significance of neither henge monuments nor the bristlecone pine is better explained as a result. A much more significant role for independent dating lies in the testing of hypotheses (Deevey, 1969). Take, for example, the mid-Holocene elm decline that is widely recorded in pollen diagrams from northwest Europe. If, as has been hypothesised, this was a consequence of prehistoric agriculture, then the fall in elm pollen values should coincide with the time of arrival of the first farming communities. Dating early Neolithic sites, on the one hand, and the appropriate section of pollen diagrams, on the other, allows the anthropogenic hypothesis to be tested. In particular, if the elm decline were found to pre-date the start of the Neolithic, then this hypothesis would be shown to be false, and our agricultural ancestors exonerated from blame (see Chapter 5 for further discussion of the elm decline).

Another reason for wanting to have precise estimates for the age of Holocene events is in order to calculate past rates of change. Some events, such as eustatically controlled sea-level rise, would have occurred at the same time everywhere across the ocean; that is, the change was a **synchronous** one. More usually, however, events begin earlier at some places than at others. The spread of a disease or the movement of a glacier snout are of this type, and they are

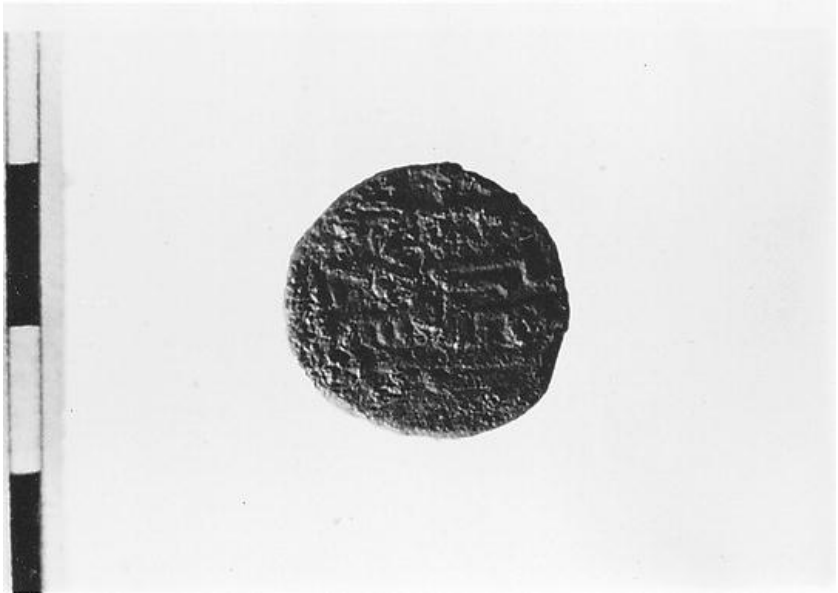
termed **time-transgressive**. The rate of disease diffusion or of glacier retreat will vary, and precise dating techniques make it possible to establish whether rates were fast or slow, constant or variable. Dating is also essential to a number of the other techniques discussed later in this chapter. Past influx of pollen or of eroded sediment into a lake, for example, can only be obtained if a sound and detailed chronology exists for the sequence of lake sediments.

One simple form of dating is provided by the fact that in undisturbed sediments, younger layers overlie older ones. This law of superposition indicates which layer was deposited first, but it fails to provide the actual age of either. Preferably, the layers should be placed not only in a relative sequence but also firmly in time, and for this it is necessary to assign them absolute ages in years. Several approaches to absolute dating will be discussed here, including those based on historical records, those employing **radiometric dating** techniques and those utilising incremental dating methods such as tree rings.

Historical and archaeological dating

Perhaps the most obvious form of historical dating is that associated with documentary-based studies of past ecology or climate. For instance, medieval manuscripts recording the extent and condition of England's royal forests almost invariably have calendar dates attached to them, as do Icelandic chronicles referring to the presence of drift ice – and hence sea surface temperature – around that island. We may go further and suggest that without an historical age, documentary records such as these are of little use in helping to piece together past environments. Dating of this kind is especially important in regions such as Europe and East Asia where there exist long historical records. Historical or archaeological evidence can also be important in dating non-documentary records such as pollen diagrams. In the case of one sediment core taken from southwest Syria, the presence of an exotic pollen type, that of maize (*Zea mays*), helped show that the upper part of the core dated to recent centuries and not to the early Holocene as had previously been proposed (Bottema, 1977). Maize is native to the Americas and was only introduced into the Old World as a crop after the Spanish conquest of Mexico in the sixteenth century.

Plate 2.1 This Arab coin from Crete, made between 847 and 861 AD, helped date the alluvial fill in which it was found.



The remnants of past human activity can provide other important clues for dating Holocene environmental changes. A Roman tile drain buried beneath 5 m of alluvium, or a classical port now silted up and many kilometres from the sea, both testify to active sediment transport and deposition by rivers during the last 2000 years. Artifacts including pottery, stone tools and coins can all be assigned ages with greater or lesser precision (see [Plate 2.1](#)), as can less obvious cultural evidence such as hemp fibre from retting that has been incorporated in lake sediments. If discovered in a stratigraphic sequence, artifacts provide maximum ages for the layer in which they were found, maximum because they may have been reworked since being deposited initially. A bicycle frame pulled from supposedly mid-Holocene coastal dunes provides a – revised – maximum age for the dunes, for even if the bicycle were manufactured 50 years ago, it could have been dumped there as recently as last year. On the other hand, structures such as a former irrigation channel or a shell midden will not have been redeposited in the same way as artifacts can be, and they may therefore offer tighter dating control. In fact, built structures will often provide minimum rather than maximum ages, say for a land surface on which they are found. However, the use of human artifacts for dating purposes can be fraught with dangers, as it necessarily involves phenomena that are time-transgressive. The Stone Age may have ended by 4000 years ago in Europe, but it arguably lives on with certain isolated hunter-gatherer groups in the tropics.

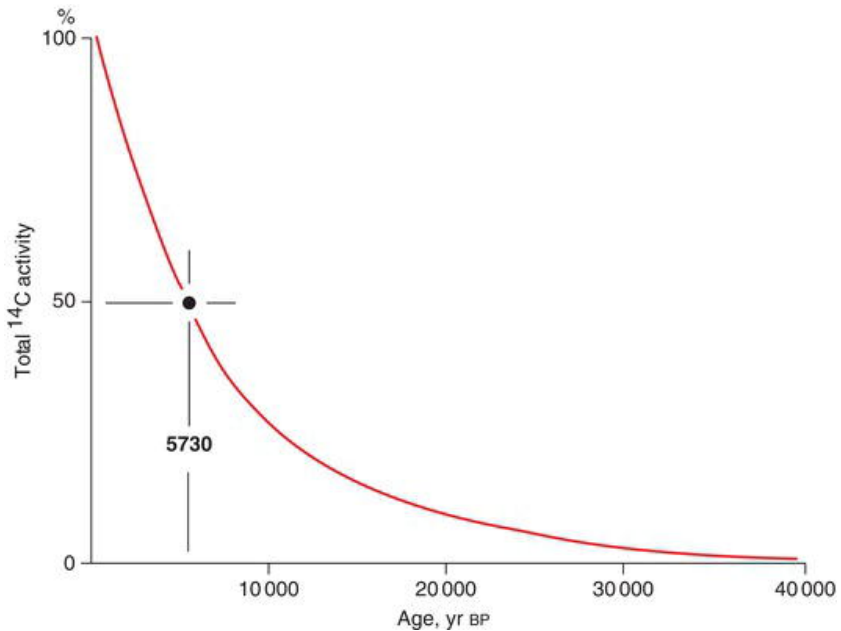
Radiometric dating methods

Radiometric dating techniques were discovered and have been applied since the middle of the twentieth century. They involve the radioactive properties of

different materials which contain within them a natural time signal, most often involving the principle of isotopic decay. Most natural elements are a mixture of several **isotopes**, which have the same chemical properties and atomic numbers but different numbers of neutrons and hence different atomic masses. One isotope is always dominant for each element; for instance, in the case of carbon, the dominant isotope is carbon-12 (or ^{12}C). Carbon isotopes with different atomic masses are ^{13}C and ^{14}C , of which the latter is by far the least abundant. Carbon-14 is also different from the other two in that it is isotopically unstable; in other words, it decays to form stable ^{14}N over time. Most importantly, unstable isotopes such as ^{14}C decay at a fixed rate. The rate of radioactive decay is not a straight line but occurs rapidly to begin with and slows down progressively over time (see **Figure 2.1**). Because decay rates are measurable, unstable isotopes represent natural archaeological or geological clocks. The clock is set to zero when isotopic decay starts, and the time elapsed can be gauged from how far the decay process has proceeded since then, as indicated by the amount of radioactivity left in the element. The decay rate for a particular isotope is most usually recorded by its half-life, or the time taken for its radioactivity to be reduced by half. Half-lives of different elements vary from the very short (e.g. radon-222, 3.8 days) to the very long (e.g. potassium-40, 1300 million years). In general, the useful dating range of individual isotopic methods is about 10 times their half-life. Beyond this, radioactive emissions are difficult to distinguish from normal background levels of radioactivity. Consequently, different isotopes provide dating methods for very different time periods, and only a few of these will be useful for the Holocene.

The most important of the radiometric dating methods for the Holocene is based on the isotope carbon-14. **Radiocarbon dating** was pioneered by Willard Libby at Chicago in the late 1940s, work which later earned him the Nobel Prize for Chemistry (Burleigh, 1981). ^{14}C is formed in the upper atmosphere by cosmic ray bombardment of nitrogen (N) atoms. The resulting carbon isotope is rapidly oxidised to form carbon dioxide (CO_2) and is then mixed uniformly and rapidly through the atmosphere. Photosynthesis leads to ^{14}C being taken up by plants, which in turn is passed on to higher organisms including ourselves. Consequently, ^{14}C is present in small but significant quantities in all living organisms. When organisms die, their ^{14}C content is no longer replenished, and the isotopic clock is set in motion. Libby calculated the half-life of ^{14}C experimentally as 5560 ± 30 years, a value close to the present best estimate of 5730 ± 40 years. Libby cross-checked samples which he had dated by the ^{14}C method against materials of known historical age, most of which came from ancient Egypt and were up to 5000 years old. The close level of agreement led him to believe that his radiocarbon dating method was a valid one and that it could be applied universally.

Figure 2.1 Decay curve for radiocarbon.



Since its discovery, many thousands of ^{14}C age determinations have been carried out by over hundred different laboratories. About half of all dates have been for archaeological investigations, with most of the rest having been for studies of past environments, and with the vast majority of radiocarbon dates being of Holocene age. The results of radiocarbon dating are usually presented as ages in years before present, normally written using the notation ^{14}C yr BP. In practice, the present is taken as AD 1950 in order to prevent dates appearing to be older simply because they were analysed more recently. Archaeological chronologies use the BC/AD system more often than the yr BP one (alternatively known as BCE/CE; Before and during the 'Common Era'), but the former can be turned into the latter by simply adding 1950 or – as a quick approximation – 2000 years to BC dates. Also listed along with each individual ^{14}C date are a laboratory code number and an error function representing one standard deviation about the mean. Thus, 8160 ± 110 ^{14}C yr BP (BM – 1666) was dated at the British Museum radiocarbon laboratory and indicates that there is a 68% probability that the ^{14}C age of this sample lies between 7940 and 8380 yr BP. It should be borne in mind that radiometrically determined dates are statistical estimates, and they should be treated as such. ^{14}C dating can be used to determine the age of any material containing carbon, including wood, charcoal, peat, seeds, bone, carbonate, shell, iron, cloth, rope, groundwater and soil. Among the more exotic dated materials have been ostrich eggshell, lime mortar and human brain!

The ^{14}C method cannot, however, be applied directly to samples of very recent age. The burning of fossil fuels since the Industrial Revolution has caused an injection of geologically old carbon into the atmosphere which has lowered its ^{14}C content and made recent samples seem older than they actually are – the so-

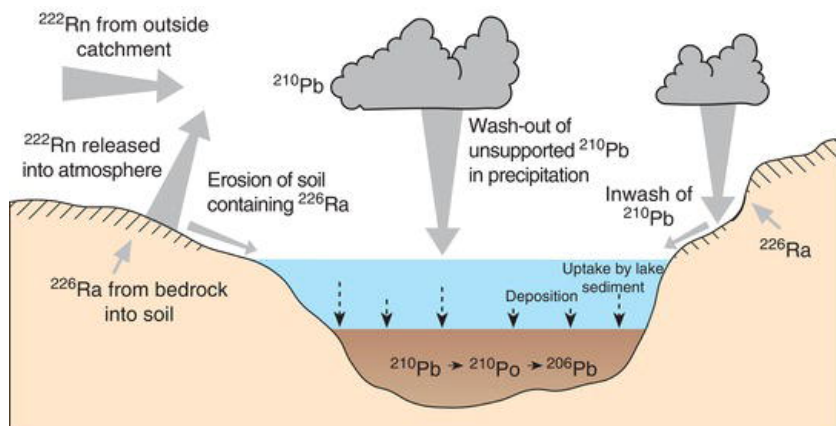
called Suess effect. More recently still, this has been overcompensated and further complicated by the testing of atomic weapons. In short, conventional ^{14}C dating is of limited use for samples younger than about 150 years. Many other forms of sample contamination are possible, and although ash from a careless investigator's cigarette has doubtless accounted for more than one erroneous ^{14}C date, contamination is usually of natural rather than of human origin. Carbon both older and younger than the date of the death of the sample may be involved. The latter is a particular problem in areas of limestone or coal-bearing bedrock, and in those experiencing active volcanic activity. This problem can be especially difficult to assess where bicarbonate-rich lake waters have been taken up by aquatic plants which subsequently formed part of the lake sediment. The dilution of ^{14}C levels causes ages to appear older than they actually are, known as the hard-water effect (Deevey et al., 1954). Contamination by young carbon, by contrast, makes dates appear too recent. There are countless such sources, including rootlets penetrating into underlying stratigraphic layers, recalcification of mollusc shells, and animal burrows in archaeological sites. However, many problems of contamination by younger material can be avoided by careful selection of samples in the field and by pretreatment during dating. Young carbon is, in any case, a much more serious problem for samples of Pleistocene than of Holocene age. Since 1980, improved dating precision has also been achieved by measuring ^{14}C directly using a mass spectrometer (Hedges, 1991). This AMS (or Accelerator Mass Spectrometer) dating method allows the dating of small samples weighing as little as 10 mg. It allows scientists to pick out and date individual seed remains, especially of land plants which should be free of any hard-water error, or charcoal fragments. More controversially, it allowed the age of threads from the Turin Shroud to be determined, which could otherwise have been dated only by destroying the shroud itself. The result showed the shroud to be a – very clever – medieval forgery, fabricated between AD 1260 and 1390.

Some materials are often considered to be more prone to contamination than others – for example, bone, shell and soil – but ^{14}C dating will be more reliable if samples of these materials are adequately prepared before age determination. In the case of bone, only the protein collagen should be dated (Gillespie, 1984, p. 13), while mollusc shell should be tested by X-ray diffraction and acid leached or mechanically cleaned if necessary. Soil, of course, takes longer to form than either of these two, but the consequent broad time range associated with ^{14}C dates on buried soils is as much a stratigraphic as a dating problem (Matthews, 1985). ^{14}C age determination of soil should ideally involve dating selected organic fractions; but even without this, Rothlisberger and Schneebeli (1979) obtained consistent and meaningful results in their study of Holocene soil and moraine stratigraphy in the Swiss Alps. All of these contamination problems are local to sites under study. By contrast, the one major revision to the ^{14}C timescale since Libby's initial discovery involves variations that were global in scale. It was initially assumed by Libby that there had been no significant changes in atmospheric levels of ^{14}C during recent millennia. Libby carried out ^{14}C determinations on historically dated materials to check this, but after initial enthusiasm, he subsequently found that older dates systematically underestimated the true age of samples. This was confirmed in 1965 when Hans Suess presented results comparing ^{14}C dates with those from another dating method – dendrochronology. These results are

discussed in detail later, but suffice it to say that ^{14}C dates older than 2500 years significantly underestimate actual age. But because these deviations from true age are systematic and worldwide, it is possible to apply correction, or calibration, factors to ^{14}C determinations to turn them into true, or calendar dates. The need for ^{14}C calibration has not, therefore, invalidated the ^{14}C method, which remains a remarkably robust and successful one.

Other radiometric techniques of increasing importance are Luminescence and Uranium–Thorium (U-Th) dating. **Luminescence**, or optical, dating was initially applied to archaeological materials such as pottery, burnt clay and flint (Aitken, 1990). It was then discovered that ‘bleaching’ of the sample, which resets the radiometric clock, could be accomplished by exposure to sunlight as well as by firing. Thermoluminescence (TL) dating subsequently came to be successfully applied to the quartz or feldspar grains of windblown sediments such as sand dunes and loess, assuming that they were not deposited in night-time darkness! Other variants have since been added to this technique, such as Optically Stimulated Luminescence (OSL), which require shorter ‘bleaching’ times and are capable of dating other types of sediment such as river alluvium (Aitken, 1994; Duller, 1996; Wintle, 2008). U-Th dating is based on a more complex decay chain than ^{14}C , with a sequence of ‘daughter’ isotopes being produced from the original ‘parent’ (Smart, 1991; Ivanovich and Harmon, 1995). On the other hand, the half-lives involved are longer than with ^{14}C , so that the technique can be used to date older materials. Unstable uranium or thorium isotopes are taken out of water by corals, or precipitated in cave speleothems or in lake carbonates. U-Th, especially with the use of a high-precision mass spectrometer, has proved valuable in calibrating the radiocarbon method beyond the range of dendrochronology (Edwards et al., 2003; Fairbanks et al., 2005).

Figure 2.2 Lead-210 pathways in a lake catchment. Adapted from Wise (1980).

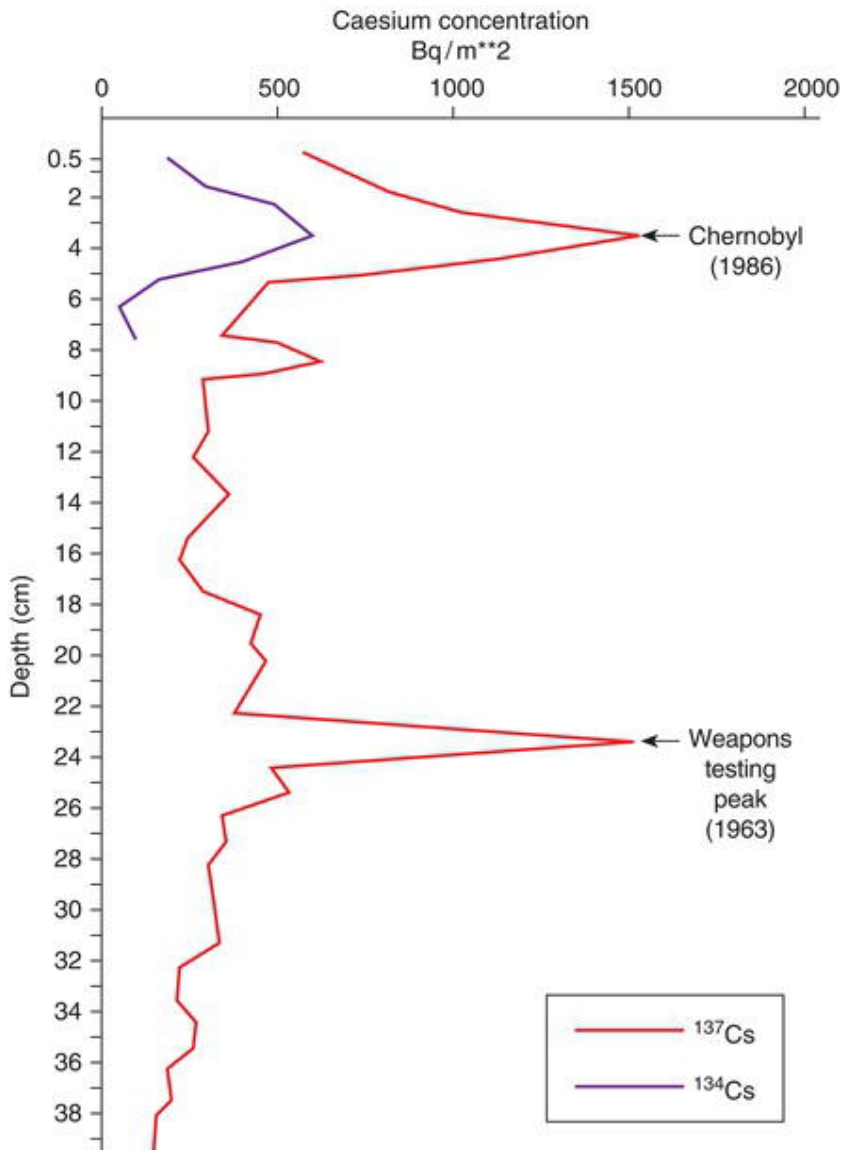


Luminescence, U-Th and ^{14}C all cover time spans of thousands of years or longer, and none is particularly useful for dating changes during the last century. Partly because of this, many palaeoenvironmental studies stop precisely at that point in time when human impact becomes most apparent. However, a number of dating techniques with short half-lives have been developed which allow the

timing of recent human impact on ecosystems to be determined. Among the most useful of these radiometric dating techniques are **lead-210** and **caesium-137**. Lead-210 (^{210}Pb) is an unstable isotope with a half-life of just over 22 years and a dating range of 100–200 years, thus conveniently complementing ^{14}C . Like U-Th, ^{210}Pb forms part of a longer decay chain which, in effect, begins when radium-226 (^{226}Ra) is released from bedrock as part of catchment erosion. Some will enter streams and lakes and will decay *in situ* to form ^{210}Pb . However, a proportion of ^{226}Ra escapes into the atmosphere as the inert gas radon-222, which decays via a series of short-lived isotopes to form ^{210}Pb . This excess or unsupported ^{210}Pb returns to Earth as rainfall or as dry fallout to augment that produced by *in situ* decay (see [Figure 2.2](#)). Only the unsupported ^{210}Pb is required for dating purposes. Various models of age estimation exist, the most comprehensive of which assumes a constant rate of supply over time (Appleby and Oldfield, 1978). In practice, however, changes in lake hydrology or catchment soil erosion rate can result in an irregular flux of unsupported ^{210}Pb (Oldfield and Appleby, 1984).

^{210}Pb and ^{14}C are naturally occurring radioisotopes, but there also exist others of artificial origin, one of which is caesium-137 (^{137}Cs). A main source of these artificial isotopes was the atmospheric testing of nuclear weapons, which started in the 1950s, reached a maximum in the late 1950s and early 1960s and subsequently declined with the advent of the nuclear test ban treaties. ^{137}Cs fallout peaked in the northern hemisphere around 1963. Another source of caesium-137 comes from atomic energy production, including the nuclear accident at Chernobyl in 1986. Investigations have shown that ^{137}Cs is strongly adsorbed onto sediment or soil and that it has been incorporated into those deposited in the last 60–70 years. The changing concentration of ^{137}Cs in the uppermost sediment layers of lakes in eastern North America and Europe parallels the monitored history of the isotope deposition on land ([Figure 2.3](#)). ^{137}Cs and other short-lived radioisotopes such as Beryllium-7 (^7Be) thus act as tracers suitable for studying soil erosion and sediment accumulation rates during recent decades (Walling and Quine, 1990; Blake et al., 2002). ^{137}Cs also provides a useful cross-check for ^{210}Pb chronologies, and a combination of the two methods provides a strong basis for determining the timing of human impact on ecosystems during the Industrial era.

Figure 2.3 Stratigraphic deposition of radioactive caesium at Ponsonby Tarn in the English Lake District (modified from Bonnett and Cambray, 1991; reproduced with permission of Springer Science + Business Media).



Dendrochronology and radiocarbon calibration

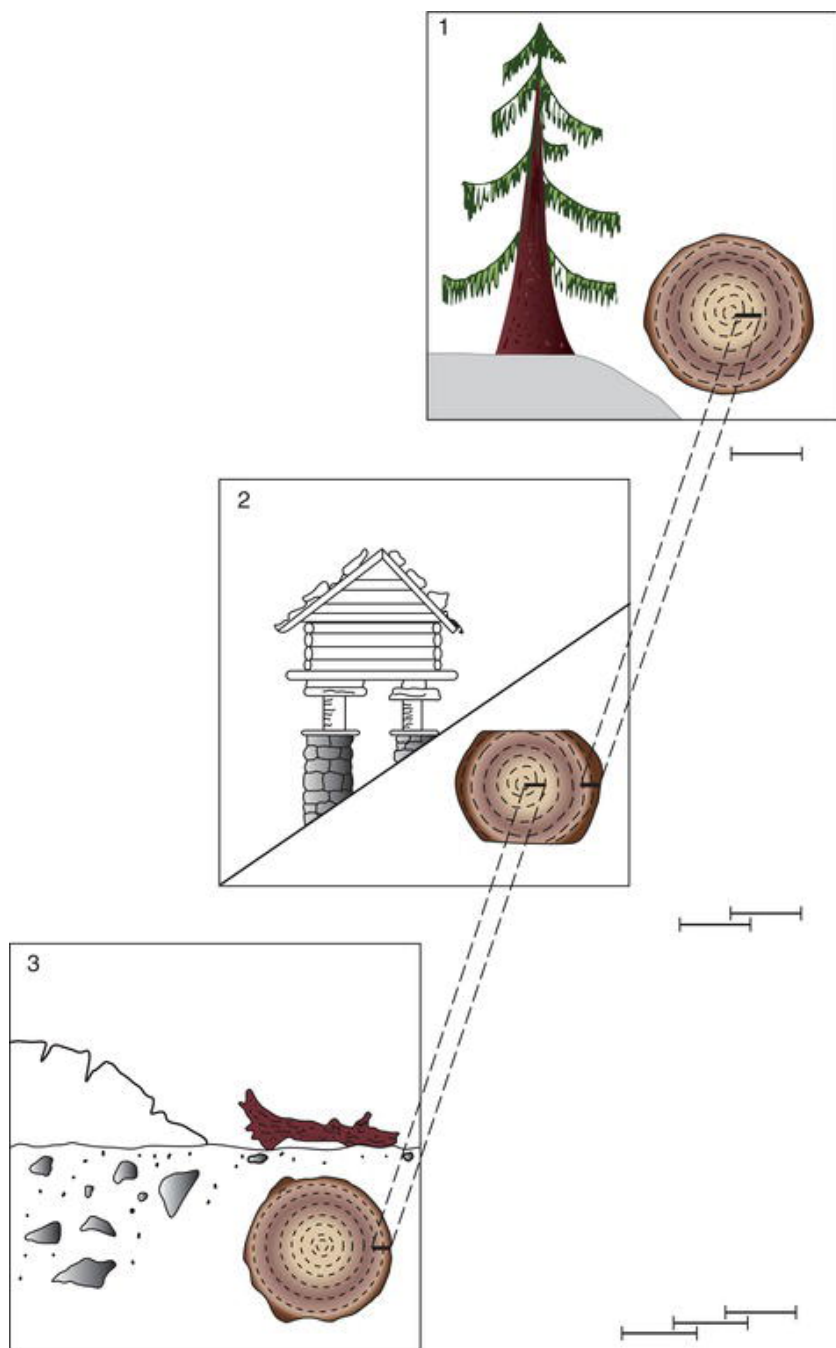
Incremental techniques, which are based on natural seasonal rhythms or annual growth rates, have been utilised since the early years of the twentieth century. However, it would be wrong to imagine that they have been made redundant by

the advent of radiometric dating techniques. As will be made clear, they remain as valuable today as they ever were.

The principles of this type of dating can be illustrated by the best-known incremental dating method: tree-ring dating or [dendrochronology](#). This technique was largely developed through the efforts of A.E. Douglass working in Arizona between 1910 and 1940. Although it had been known since at least the time of Aristotle that trees produced annual growth rings, Douglass was the first to systematise this into a dating method. At its simplest, the technique involves counting the number of growth rings present between the bark and the centre of a living tree. As the outermost ring can be assumed to represent the current year's growth, the age of the tree can be calculated by counting the number of rings present. This need not require lumberjack skills, however, for a tree-ring sequence can be obtained without harming the tree by using an increment borer, which provides pencil-sized tree cores. Tree-ring dating can be used to provide a chronology for valley glacier retreat or alluvial fan activity over recent decades and centuries, although the ages it provides in these cases are only minimum ones because of the time lag between deglaciation or channel change and subsequent colonisation by trees. Potentially of greater value are those approaches which exploit the characteristic tree-ring signatures produced as a result of differential growth from one year to the next, which in turn are linked to variations in the weather (see [Technical Box X](#)).

Tree-ring signatures make it possible to cross-date or correlate separate sequences. Such cross-dating is not restricted to live wood, so that sequences from dead trees or wood used in buildings can be matched up with those from living trees (see [Figure 2.4](#)). The former would otherwise be free-floating chronologically, as their outer edges do not date to the present day and consequently lack a firm reference point in time. The building up of a chronology from separate tree-ring sequences amounts to no less than a jigsaw puzzle with time. 'The pieces are scattered round as living trees, stumps, timbers in buildings and buried, either as archaeological material or as naturally preserved timbers, in bogs, rivers or lake beds' (Baillie, 1995). In Arizona, Douglass compiled both a master chronology stretching from living yellow (ponderosa) pines back to the thirteenth century AD and a floating one based on pine timbers from prehistoric buildings. In 1929, after much searching, he found the missing piece in his jigsaw – an excavated wooden beam from Pueblo Bonito dating to between AD 1237 and 1380 – and the two sequences were joined.

Figure 2.4 Cross-dating alpine timbers from overlapping tree-ring signatures.



Huber subsequently applied Douglass's approach to German tree rings, but based on oak rather than pine timbers. This oak chronology has since been

extended and joined to a ‘floating’ pine sequence which together cover the whole Holocene and go back into the preceding late-glacial period (Friedrich et al., 2004). A similar 7000+ year chronology has built up from Irish bog oaks using the same cross-dating approach. The most famous dendrochronological record, however, comes from the bristlecone pines (*Pinus longaeva* and *Pinus aristata*) (Plate 2.2), which grow at altitudes of 3000 m in the mountains of Southwest United States. The Great Basin bristlecone pine is the world’s longest living tree, and incredibly, living specimens have been recorded up to 4700 years old, with the oldest known example being named Methuselah. Cross-dating of sub-fossil trunks with living trees has produced a continuous bristlecone pine tree-ring chronology over 8000 years long. The construction of tree-ring chronologies such as these involves the science of computer matching and X-ray densitometry as well as the art of matching tree-ring series by eye (Baillie, 1995; Stokes and Smiley, 1996).

One of the most important features of dendrochronology is the fact that the wood used for tree-ring counts can also be dated by radiocarbon, and this has allowed a cross-check to be made between the two methods. At first sight, it may seem surprising that it is possible to get ‘old’ dates from ‘living’ wood, but most of the woody cells in trees are empty, being used merely for transporting water from the roots to the branches. The cells laid down in each growth ring are the only part to retain a new ^{14}C imprint. In the 1960s, Hans Suess radiocarbon dated samples of bristlecone pine of known tree-ring age. His results revealed a systematic divergence between ^{14}C and tree-ring ages in samples more than about 2500 years old (see Figure 2.5). Mid-Holocene ^{14}C dates appear to be too young, with a discrepancy amounting to 600 years (or 12%) for material 5000 years old. Tree-ring dates are more accurate than their ^{14}C equivalents, although this does not invalidate the ^{14}C method but instead requires that ^{14}C dates be calibrated if they are to be expressed as true, or calendar years. Tree rings have been joined by other dating methods which can be paired against radiocarbon, such as U-Th ages from corals and varved lake sediments, which extend ^{14}C calibration back to the time of the last glaciation (Fairbanks et al., 2005; Bronk Ramsey et al., 2012).

There is every reason, therefore, to adopt a calendar timescale for an environmental history of the Holocene. In consequence, **the ages in this book will use the convention of quoting calibrated ^{14}C dates, expressed as Cal. yr BP** (calendar years before present) or, as a shorthand version, **ka BP** (thousand years before present, with k = kilo, a = annum). Calibration has been based on INTCAL (Reimer et al., 2004 and online updates) via OxCal. A summary conversion table from ^{14}C to calendar ages (and vice versa) and weblinks can found in the appendix. In this book, calibration has been applied to both individual ^{14}C dates and overall chronologies. In some published papers, ^{14}C determinations are quoted only as uncalibrated dates, and they need to be converted to calendar ages in order to make a comparison with the timescale used here. It is important to recognize that there are some potential pitfalls in making a conversion from ^{14}C to calendar years. Firstly, the tree-ring calibration curve is not a smooth line but contains a number of wiggles which represent short-term variations in atmospheric ^{14}C content (Figure 2.5). This is significant in that some ^{14}C dates have more than one calendar age; a ^{14}C date of 4500 yr BP for instance has calibrated ages spanning almost 200 years between 5056 and 5246 years.

Calibration can therefore have an important effect on studies where high dating resolution is required because an apparently synchronous event may be turned into one that is potentially time-transgressive. This is a particular problem for dating the onset of the Holocene ([Technical Box I](#)). In high-resolution studies, on the other hand, it may be possible to obtain sufficient ^{14}C dates to use the ‘wiggles’ of the calibration curve to match up the series being dated. A second problem is that ^{14}C calibration beyond about 13 ka BP is not yet universally agreed. Nonetheless, all techniques agree that ^{14}C and calendar ages continue to diverge back to the [Last Glacial Maximum \(LGM\)](#). For our purposes, they cross-match sufficiently well to make it possible to use a calibrated timescale for all of the events and changes covered in this book, even if the precision achieved for ^{14}C calibration is somewhat reduced in late Pleistocene times.

Plate 2.2 The world’s longest living tree, the Bristlecone pine (*P. longaeva*), which lives at high altitude in the White Mountains of California.

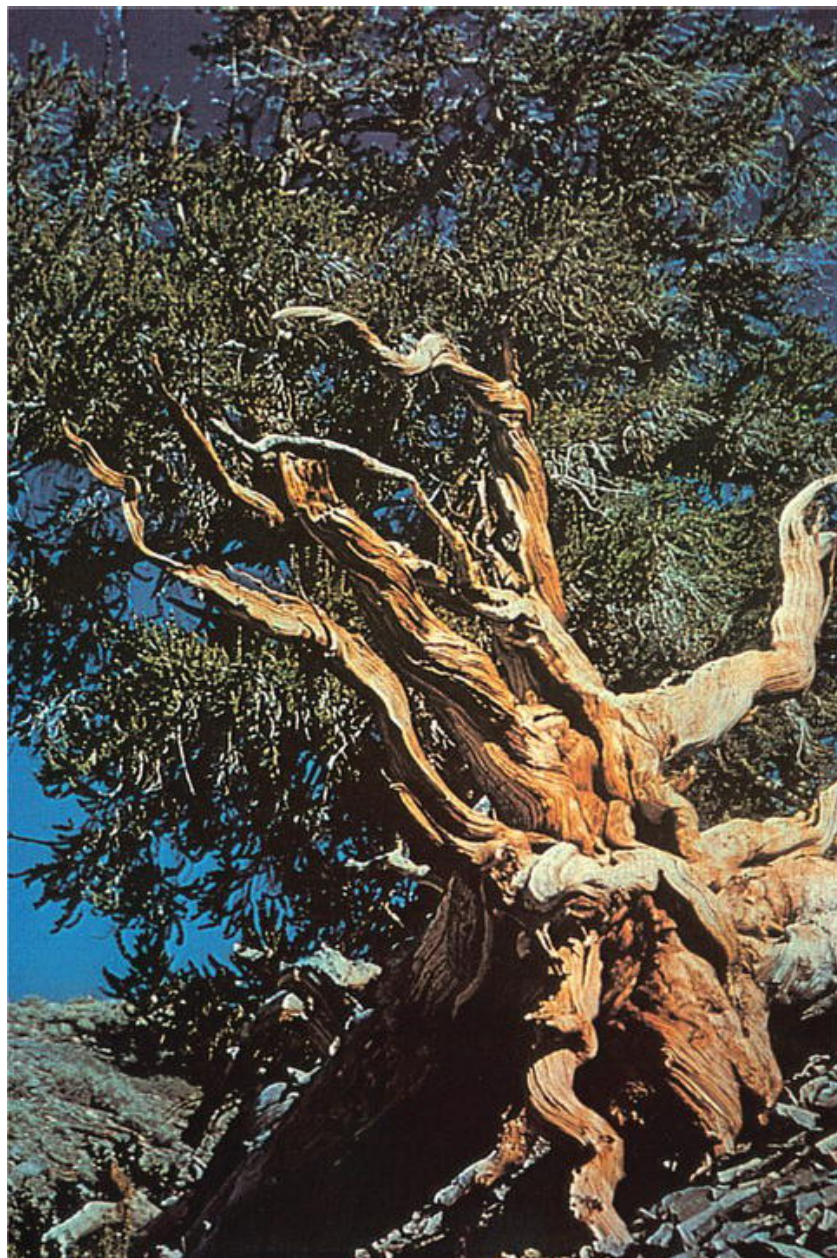
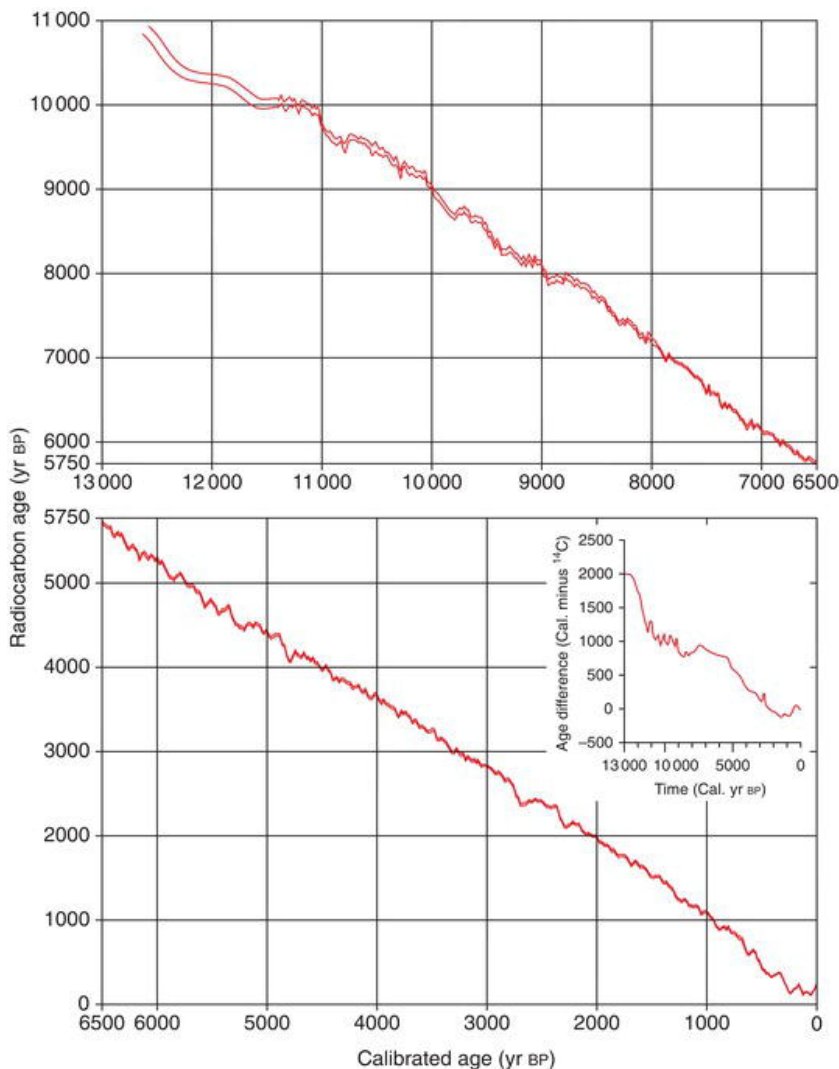


Figure 2.5 Tree-ring radiocarbon calibration curve for the Holocene, from OxCal program (inset: departures from calendar age with time); Younger Dryas based on Goslar et al. (1995).



TECHNICAL BOX

Technical Box I: Dating the onset of the Holocene

There are different schools of thought about how the start of the Holocene should be formally defined. The first school of thought (e.g. Watson and Wright, 1980)

believes that the beginning of Holocene is easily recognised in individual records of deglaciation or biotic change (e.g. in pollen diagrams), but that because these changes are not exactly the same age everywhere, the Pleistocene-Holocene boundary should therefore be time-transgressive. A second school prefers the 'type-section' approach commonly used in geology to define stratigraphic boundaries. Walker et al. (2009), for example, proposed that the NGRIP Greenland ice core be used as the standard reference point for the Pleistocene-Holocene boundary. Annual ice layer counting of the NGRIP core provides an age of $11\,700 \pm 100$ calendar yr b2k (before AD 2000) for the sharp change in climate proxies marking the start of the Holocene. A third view is that the Holocene should be simply defined as beginning 10 000 ^{14}C years ago. Radiocarbon, after all, provides the most common way in which we date individual climatic and archaeological record, but the difficulty here is in knowing what 10 000 ^{14}C yr BP means in terms of real calendar ages. ^{14}C -dated tree-ring sequences have shown a systematic divergence between the two timescales which amounts to thousand years or more by the beginning of the Holocene. Moreover, ^{14}C time is elastic; that is, it has been stretched or compressed at different periods in the past. And perversely – if not coincidentally – 10 000 ^{14}C yr BP happens to coincide with a major ^{14}C 'plateau'; when the ^{14}C 'clock' stood still, while at least 400 years of real time elapsed (Taylor et al., 1996) (Figure 2.5). This ^{14}C plateau may have been caused by abrupt changes in atmospheric CO_2 and oceanic circulation at the end of the last glaciation (Muscheler et al., 2007), and it has meant that 10 000 ^{14}C yr BP represents not one moment in time, but several! It would clearly be preferable to define the onset of the Holocene in calendar and not in ^{14}C

years.

So which date should be chosen? There is evidence of very rapid climatic change in a number of independently-dated natural archives, and the different calculated ages for this change are shown in [Table 2.1](#). These different techniques provide similar results, with most of them lying between 11 800 and 11 300 Cal. yr BP. The INTCAL09 data set gives $11\,500 \pm 100$ Cal. yr BP as the calibrated equivalent of 10 000 ^{14}C yr BP (Reimer et al., 2004), although this is about 200 years younger than counts of annual snow layers in Greenland ice cores (Alley et al., 1993; Rasmussen et al., 2006). The Sub-commission on Quaternary Stratigraphy recommended that the NGRIP ice core be used as the global stratotype section for the base of the Holocene Epoch, and this was subsequently agreed by the International Union of Geological Sciences. On that basis, an age of 11 700 Cal. yr BP is now conventionally used for the start of the Holocene, although it is recognised that this age could be in error by a century or more either way.

Table 2.1 Calendar age estimates for the beginning of the Holocene

Sources: Alley et al. (1993), Björck et al. (1996), Gulliksen et al. (1998), Friedrich et al. (2004), Rasmussen et al. (2006) and Walker et al. (2009)

Age (Cal. yr bp)	Method
11 590	Calibration of ^{14}C dates on Central European oak + pine tree-ring chronology
11 360–11 920	Counting of annual layers in GISP2 Greenland ice core
11 600–11 800	Counting of annual layers in GRIP and NGRIP Greenland ice cores
11 200–11 700	U-Th calibration of ^{14}C in tropical corals
11 546–11 638	Varve calibration of ^{14}C , Cariaco marine basin,

11 395–11 540	Caribbean Sea 'Wiggle matched' ^{14}C lake sediment record, Kråkenes Lake, Norway
11 360–11 600	Varve calibration of ^{14}C , L. Gosciaz, Poland
c.11 600	Varve calibration of ^{14}C , Holzmaar, Germany
11 440	Varve counting, Sweden
11 464–11 640	Varve calibration of ^{14}C , Lake Suigetsu, Japan

Other dating methods

Dendrochronology is only one of a number of incremental dating methods. Another technique with high accuracy and precision involves counting the annual layers of snow which build up to form ice sheets. If the accumulation significantly exceeds the summer melting, and if it is not deformed by lateral movement (as it is in a valley glacier), then the snow layers will be preserved even after later compaction turns them to ice (see [Plate 2.8](#)). After a certain point, these layers become so squashed by the overlying weight of ice that they are no longer visible. Nonetheless, in the Greenland Ice Sheet Project (GISP2) core, it was possible to count every annual layer back more than 17 000 years with an estimated precision of $\pm 3\%$ (Alley et al., 1993).

In similar fashion, some lake and marine sediments may be finely layered, or laminated, representing one year's accumulation. Annually laminated lake sediments are known as [varves](#) and can form in a number of different ways. The varves studied by the Swedish scientist Baron de Geer at the end of the nineteenth century had formed in lakes adjacent to the former Scandinavian ice sheet and comprised couplets of alternating coarse- (summer) and fine-grained (winter) layers (Sturm, 1979). These clastic varves provided one of the first quantitative estimates for the duration of the Holocene. On the other hand, laminated lake sediments can form wherever there is a strong seasonal variation in sediment supply or where circulation of water is absent or incomplete for at least part of the year (O'Sullivan, 1983; Zolitschka, 2003). One of the commonest situations in which non-glacial varves form is in lakes which freeze over in winter, as is the case in much of North America and Scandinavia. Organic matter accumulates in the still-water conditions beneath the ice cover, often followed in spring by a deposition of [allochthonous](#) sediment washed into the lake during snowmelt. In the summer months, photosynthesis by algae may lead to a layer of [diatoms](#) settling out on the lake bed. [Anoxic](#) conditions suitable for varve formation can also occur in some marine basins whose waters are incompletely mixed with those

of the open ocean, for example, off the coast from Santa Barbara in southern California. Varves can also be chemical in origin if lake water is either strongly acid or alkaline, or if it becomes saline. Alkaline lakes such as Zürichsee in Switzerland are rich in carbonate which summer photosynthesis causes to be precipitated out as a CaCO_3 white-out (Kelts and Hsü, 1978). Lake Suigetsu on Honshu Island, Japan, has been depositing varved sediments for more than 50 000 years which, coupled to almost a thousand ^{14}C dates, has allowed calibration of the radiocarbon timescale back into late Pleistocene times (Bronk Ramsey et al., 2012; Nakagawa et al., 2012). Varves have proved useful even when they are not continuous to the present day and are therefore free-floating in time, for example, in calculating the length of the cold Younger Dryas stadial at the end of the Last Ice Age (Lotter, 1991).

Using **lichens** for dating achieves a very different degree of precision. Lichens represent a symbiotic relationship between fungi and algae. The algae photosynthesise to provide nutrition for the partnership, and the resulting compound organism behaves as if it were a single biological unit. Lichens are able to exploit a wide range of habitats and grow on substrates ranging from tree trunks to bare rocks. Because of their ability to survive in unfavourable environments, they often take the role of primary colonisers in plant succession. Their utilisation for dating purposes exploits the fact that most lichen species grow at rates that are both fixed and measurable. The larger the lichen, the greater its age. Empirical work by R.E. Beschel in arctic and alpine environments during the 1950s and 1960s showed there to be a clear relationship between lichen size – as measured by diameter – and age. Beschel went on to formalise this as a dating technique which has subsequently been applied to many field studies, notably of Holocene glacier fluctuations (Innes, 1985). Lichenometry is especially valuable for establishing a date for exposure of a bare rock surface following deglaciation (Matthews, 1992). Field techniques for the measurement and sampling of lichens for dating are relatively straightforward (Lock et al., 1979), but doubts have been raised about whether the assumptions behind them have been adequately tested (McCarroll, 1994). In particular, it remains uncertain how far lichen growth rates are affected by variables other than time, including competition for space, light and nutrients. In consequence, lichenometry is best considered as a dating method that can provide relative rather than absolute ages.

Another means of dating involves the fact that the Earth acts as a giant magnet with its own magnetic field, and that this field changes constantly. Perhaps the most dramatic example of these changes is the periodic reversals in the Earth's magnetic polarity. However, the last time the compass needle slipped from the South to the North Pole was over 700 000 years ago, and although there have more recent short-lived magnetic excursions, polarity reversals occur too infrequently to help date the Holocene. On the other hand, so-called secular magnetic variations can be important in helping to provide a timescale over hundreds or thousands of years. It has long been recognised that magnetic north does not coincide with geographic or true north and that the position of the former has wandered through time. Measurements of the geomagnetic field have been taken at London for over 400 years and show that magnetic north migrated westwards between AD 1580 and 1820, when it lay 20°W of true north. Since then, it has shifted eastwards once more.

The angle between true and magnetic north, or declination, is only one of three components of the magnetic field. The compass needle will not only move from side to side but, if freely suspended, dip down at an angle towards the Earth. This dip or inclination varies from the equator to the poles, as does strength or intensity of the Earth's magnetic field. All three of these components vary through time and in combination produce a characteristic palaeomagnetic signature. Their magnetic alignments can be incorporated and preserved in baked materials or in sediment particles which settle out in standing water. The study of archaeomagnetic samples such as hearths has extended the record of secular magnetic variations back to before AD 1000 (Aitken, 1990). For earlier periods, magnetic measurements have been applied to wet lake sediment cores with some success (Thompson and Oldfield, 1986). Secular magnetic variations do not, on their own, provide age estimates. However, once a master geomagnetic curve has been dated by another method such as ^{14}C , it can be applied to sites elsewhere. Holocene master curves, which are regional rather than global in extent, now exist for several areas of the world (Thompson, 1984). Measurement by magnetometer is rapid and, in the case of declination, non-destructive. On the other hand, not all lakes are equally well suited for the technique, which is effectively restricted to lake sediments with homogeneous particle size and organic content down core.

Conclusion

Possibly the most notable feature of this review of dating techniques is the great diversity of ways in which age estimates can be obtained for the Holocene. Dating techniques also differ in the age ranges over which they can operate, in the sample materials they use, and whether they destroy the sample during dating (see Table 2.2). Overall dating reliability is consequently improved by the use of several techniques in combination. Some of the most precise dates, other than historical ones, come from tree-rings, lake varves and ice layers. These are all potentially accurate, can achieve a precision of one year or less, and have the additional advantage of dating materials which are prime sources of palaeoenvironmental information. On the other hand, none of these dating methods is as widely applicable as ^{14}C dating.

Dating is not only a matter of precision and accuracy but also of attribution and interpretation. Dated samples almost always come from a stratigraphic context of one kind or another, and after age determination, they must be related back to the site stratigraphy. Because methods like ^{14}C and Luminescence provide spot dates rather than a continuous chronology, the age of intervening stratigraphic layers needs to be assessed by interpolation between, or extrapolation beyond, dates obtained by these methods. This is normally achieved by use of an age–depth curve, in which dates are related to their stratigraphic position in a sediment core or profile. In sediments which are still accumulating, the uppermost layers can be assumed to date to the present; otherwise, at least two dates are necessary for an age–depth curve to be drawn up. Some computer-based age–depth models calculate statistical probabilities for the age of different

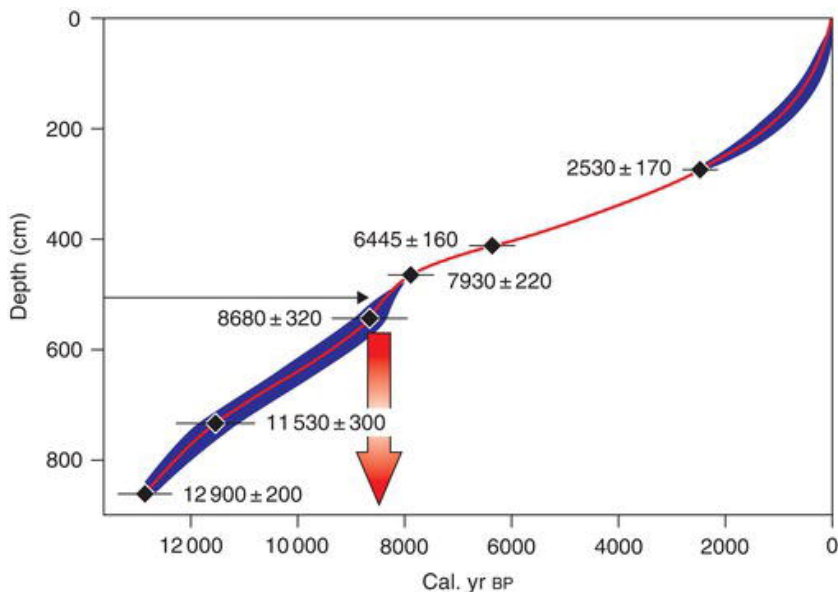
sample depths (e.g. programs such as CLAM developed by Martin Blaauw; see <http://chrono.qub.ac.uk/blaauw/wiggles/>). On the age–depth graph shown in Figure 2.6, for example, a sample from a depth of 520 cm would have a calculated age between ~8600 and ~8200 Cal. yr BP.

If change through time at a single sequence is investigated via stratigraphy, the matching of different sequences over space is achieved via **correlation**. Features such as the European elm decline are best demonstrated to be of the same age if they are dated independently at every site. Correlation using pollen or other biostratigraphic evidence is likely to be imprecise and may lead to circular reasoning (Blaauw, 2012), but other means of correlation can be a reliable alternative to independent dating. Prominent among these is the use of volcanic ash layers, or **tephrochronology**, as marker horizons (**Technical Box II**). Fossil soils can also sometimes be traced over considerable distances and are normally age-equivalent rather than time-transgressive.

Table 2.2 Dating methods applicable to the Holocene

	Time range Cal. yr BP	Precision year	Applications
<i>Historical</i>			
Archival	1–5000	<1–50	Varied; for example, glacier histories
Archaeological	50–> 40 000	1–1000	Varied; for example, alluvial histories
<i>Radiometric</i>			
Radiocarbon	200–40 000	20–1000	Organic materials
Luminescence	50–> 40 000	50–5000	Burnt clay and pottery, most minerogenic sediments (e.g. loess)
Uranium series	1000–> 40 000	20–5000	Carbonates (e.g. coral), volcanics
Lead-210	10–200	1–10	Lake and coastal sediments, peat
Caesium-137	1–35	1–5	Lake sediments
<i>Incremental</i>			
Tree ring	1–11 000	1–5	Wood timber
Varves	1–20 000	1–10	Some lake and marine sediments
Annual snow/ice layers	1–> 20 000	1–400	Ice sheet cores
Lichenometry	10–5000	10–1000	Glacial histories
<i>Palaeomagnetism</i>			
Secular variations	100–15 000	10–100	Lake sediments

Figure 2.6 Age–depth curve for a sediment core with a series of calibrated ¹⁴C dates. The best fit line and uncertainty (dark shading) have been calculated using the program CLAM. As an example, a sample from a depth of 520 cm would have a calculated age between ~8600 and ~8200 Cal. yr BP (shown by vertical arrow).



TECHNICAL BOX

Technical Box II: Tephrochronology

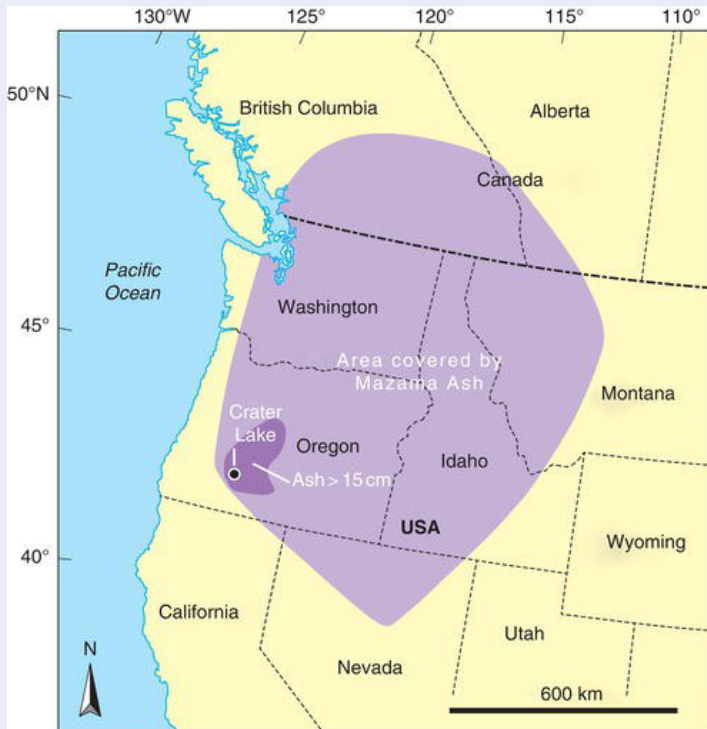
Large explosive volcanic eruptions can throw substantial quantities of fine grained ash – or tephra – into the atmosphere, where it is carried downwind, eventually to be deposited back to Earth as a layer of dust. Close to the volcano, proximal volcanic tephra layers can be centimetres or even metres in thickness; they are generally well-preserved and clearly visible in stratigraphic sequences. Because a tephra horizon from a single eruption event formed at the same time, it must be synchronous everywhere. For this reason, if a tephra can be dated (e.g. by ^{14}C) at one site, then this age can be transferred across to other sequences where it is found. This saves money on dating, but more importantly allows very precise time correlation between sequences, in some cases over large areas. One of the largest explosive eruptions of the Holocene occurred 7700 years ago in Oregon when the strato-

volcano Mount Mazama blew itself apart (Zdanowicz et al., 1999). This not only left behind the deepest lake in the US (Crater Lake, [Plate 2.3](#)) but also deposited an ash layer which can be used to correlate records across a large part of the Pacific Northwest ([Figure 2.7](#)). One key assumption in tephrochronology is that an ash layer found in different sequences derives from the same eruption. In order to confirm this, it is important to analyse the geochemical composition of individual tephra layers using major elements like Si and Fe, and – if necessary – trace elements too (e.g. Eastwood et al., 1999).

Plate 2.3 Aerial view of Crater lake, Oregon, created following the explosive eruption of Mount Mazama during the early Holocene. By Zainubrazvi (own work) [GFDL (<http://www.gnu.org/copyleft/fdl.html>), CC-BY-SA-3.0 (<http://creativecommons.org/licenses/by-sa/3.0/>) or CC-BY-SA-2.5-2.0-1.0 (<http://creativecommons.org/licenses/by-sa/2.5-2.0-1.0>)], via Wikimedia Commons.



Figure 2.7 The distribution of volcanic ash across western North America from the eruption of Mount Mazama 7700 years ago, leaving behind Oregon Crater Lake.



Not surprisingly, much research using tephra as dating horizons has taken place in regions which have been volcanically active during the Holocene, such as New Zealand (e.g. Newnham et al., 1998) and southern Italy (Zanchetta et al., 2011). However, in recent years, it has also been recognised that volcanic glass shards can be found far from the source eruption as a result of long-distance transport in the atmosphere. These so-called ‘crypto-tephra’ are not visible to the naked eye, but can be identified using microscopy. Although hunting for volcanic shards in sediment cores can involve many hours of tedious laboratory work, the rewards have been considerable. For example, crypto-tephra from Iceland’s frequent eruptions have now been found in peat bog sequences across northwest Europe (Hall and Pilcher, 2002; Lawson et al., 2012). These have led to improvements

in both regional chronologies and in correlation with the Greenland ice core climatic record, via sulphate peaks and ash shards found in the ice (Zielinski et al., 1994; Turney et al., 2004).

Palaeoecological techniques

Palaeoecology is the study of fossil organisms in order to reconstruct past environments. Accurate palaeoenvironmental reconstruction also depends on modern ecological data and is consequently only as good as our knowledge of present-day ecosystems. A fundamental assumption made here is that relationships which can be observed at the present time also held good in the past. This principle is known as **uniformitarianism** (or more correctly, actualism). Take the case of *Hippopotamus* bones found in the middle of the arid Saharan desert and dated to between 10 000 and 6000 years ago. Assuming that the bones have not been moved significantly since death, then two possible explanations might be offered. The first and more obvious would be to employ our knowledge of modern hippo ecology to infer that the Sahara was a land of lakes and rivers during the early Holocene. The alternative explanation would be that hippos have changed their habits and habitats, formerly scaling sand dunes and only recently taking to wallowing in mud!

In this example, the case for the present not being the key to the past seems somewhat absurd. But for organisms with short life cycles, such as beetles (Coleoptera), the validity of the principle of uniformitarianism may be called into question. Up to 20 000 generations of beetles have lived and died since the **LGM**, perhaps sufficient for evolutionary changes to have taken place. In fact, detailed comparisons of modern and Ice Age beetle skeletons have revealed no discernible changes in their anatomy, indicating evolutionary stability (Coope, 1970). Even so, uniformitarianism remains assumed not proven, and the further back in time we go, the more uncertain this assumption becomes. Uniformitarianism remains an important issue in other ways too (see [Table 2.3](#)), for example, in the problem of interpreting an assemblage of fossils which has no modern equivalent, or analogue (Jackson and Williams, 2004). The process by which a group of living plants or animals is transformed into an assemblage of fossils is by no means always a straightforward one. This transformation and the subsequent interpretation of the fossil assemblage are illustrated here with reference to pollen analysis. Pollen analysis, or **palynology**, is broadly representative of palaeoecology as a whole in terms of the methods it employs.

Table 2.3 Uniformitarian^a assumptions in palaeoecology

1. We understand the environmental factors governing present-day plant and animal distributions

2. Their ecological affinities have not changed through time
3. Present (and past) distributions are (and were) in equilibrium
4. Modern analogues exist
5. The origin (taphonomy) of a fossil assemblage can be established, and that...
6. This assemblage is not biased by contamination or differential preservation
7. Fossils can be identified to a meaningful level of taxonomic resolution

^a*sensu* Actualism

Pollen analysis

Palynology is the single most important branch of terrestrial palaeoecology for the late Pleistocene and Holocene. It has been extremely widely used since it was first applied in stratigraphic studies by the Swede Lennart von Post at the start of the twentieth century (Faegri et al., 1989). Its basis is that pollen grains and spores produced as part of plant reproduction are incorporated and preserved in lake muds, peat bogs or other sediments which can later be analysed to reconstruct the vegetation of an area. Pollen production starts with a life assemblage, such as mixed temperate woodland. Within the woodland ecosystem, pollen is produced more abundantly by some plants than by others (see [Table 2.4](#)); in particular, productivity is much greater for wind-pollinated than for insect-pollinated taxa. Of the many pollen grains produced, only a few are used for reproductive purposes, with the remainder being dispersed through the environment. Pollen dispersal involves several pathways, notably via high- and low-level airborne transport and via a waterborne route in runoff and streams (Tauber, 1965). Each pathway introduces its own selective bias into the dispersal process. High-level wind dispersal, for example, leads to a well-mixed pollen 'rain' with the pollen of some taxa being carried over long distances, notably the conifers (gymnosperms), some of whose pollen grains have sacs to help keep them airborne.

Table 2.4 Relative pollen productivity of selected temperate-zone plants

Source: Data were derived from Broström et al. (2008)

	Deciduous trees	Coniferous trees	Shrubs and heath	Grasses and herbs
x1 (± 0.3)	Lime (<i>Tilia</i>) Ash (<i>Fraxinus</i>) Elm (<i>Ulmus</i>)		Ling heather (<i>Calluna vulgaris</i>) Willow (<i>Salix</i>)	Sedge (Cyperaceae) Ribwort plantain (<i>Plantago lanceolata</i>) Grasses (Poaceae/ Gramineae)
x2 (± 0.5)	Beech (<i>Fagus</i>)	Spruce (<i>Picea</i>)	Hazel (<i>Corylus</i>) Juniper (<i>Juniperus</i>)	Cereal-type (<i>Cerealia</i>) Dock (<i>Rumex</i>)
x4 (± 1)	Birch (<i>Betula</i>) Hornbeam (<i>Carpinus</i>)		Wormwood (<i>Artemisia</i>)	
x7 (± 2)	Oak (<i>Quercus</i>) Alder (<i>Alnus</i>)	Pine (<i>Pinus</i>) Fir (<i>Abies</i>)		

The processes of pollen production, dispersal and deposition thus generate a death assemblage which is significantly different in its composition from the initial life assemblage. This is further altered by sediment diagenesis and by differential fossil preservation and destruction (Birks and Birks, 1980; Chapter 1). Pollen is best preserved under [anaerobic](#), typically acid conditions, as are encountered in blanket peat bogs. The by-now fossil assemblage can be investigated through field and laboratory work, typically involving coring at a field site (see [Plate 2.4](#)) followed by laboratory preparation of samples. In order that only the sporopollenin remains intact, inorganic components of the sediment such as carbonate and silica are removed by reagents like hydrochloric and hydrofluoric acids. After mounting on a glass slide, pollen grains can be examined under the microscope, usually at a magnification of 400 $\times$. Individual pollen taxa have different shapes, sizes, number of apertures and other features which allow them to be differentiated (Moore et al., 1991) ([Plate 2.5](#)). Identification is normally to the level of genus for trees and shrubs and only to family for many herbs and grasses: species-level identification is the exception rather than the rule in palynology.

We can illustrate the sequence described earlier by means of the example shown in [Figure 2.8](#). In this hypothetical case, only four plant taxa are present; pine (*Pinus*) and deciduous oak (*Quercus*) form separate woodland stands either side of a lake, holly (*Ilex aquifolium*) is an understorey shrub and there is a fringing mat of sedges (Cyperaceae) around the lake. Although pine and oak produce abundant pollen, the prevailing wind direction causes pine to be more strongly represented in the pollen rain reaching the lake. In contrast, holly is hardly represented at all, because it is insect-pollinated. The pattern of vegetation reconstructed from fossil pollen includes aspects of the original flora, but these have been modified and are incomplete. Not only does pine appear more abundant than oak, but it is difficult to know if the two types of tree formed separate or mixed woodland stands. The sedge pollen could also be interpreted in

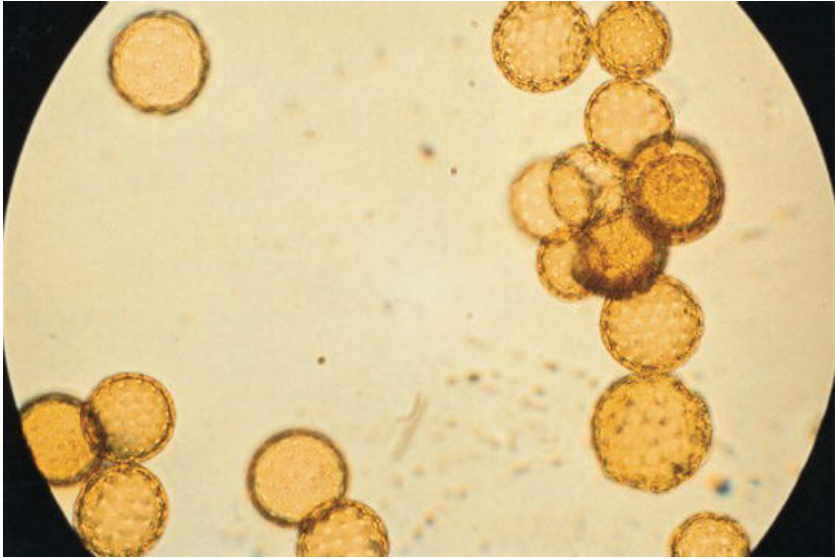
more than one way, for while some species form sedge swamps, other species occupy very different habitats such as tundra. The rarity of holly pollen grains might cause it to be overlooked altogether in the reconstruction. It is precisely in order to overcome some of the biases illustrated here that palaeoecologists carry out studies of modern pollen deposition and preservation as an integral part of their work.

The reconstruction shown in [Figure 2.8](#) is based on the relative proportions of different pollen taxa. To obtain representative proportions, it is normally necessary to count at least 200 pollen grains in any sample. However, even if a much larger number than this were to be counted, the ‘relative’ approach has certain inherent limitations. Imagine that the lake in our example was progressively infilled with sediment, and that as it became shallower, so sedge swamp largely replaced open water. Tree pollen would then be outnumbered by a superabundance of locally produced sedge pollen (see [Table 2.5](#)). In a pollen count based on relative proportions, pine and oak would appear to decline in importance over time even though the surrounding woodland vegetation was in fact unchanged. In an attempt to overcome this difficulty, palynologists often exclude some local, wetland pollen types from the total pollen sum on which percentages are based. On the other hand, it may be difficult to assign particular pollen types to a local wetland as opposed to a ‘dryland’ origin with certainty. An alternative approach is to count ‘absolute’ numbers of microfossils to obtain pollen concentration per cubic centimetre of sediment. This is most often done by adding a known quantity of exotic markers, such as *Lycopodium* spores or polystyrene microspheres, into the sample during preparation. If a timescale is available for the site under study, concentration values can be translated into rates of pollen influx per year.

Plate 2.4 Coring at the edge of Lake Gölhisar in southern Turkey. The resulting 8.3-m-long core produced a pollen and stable isotope record spanning most of the Holocene, along with a volcanic tephra layer from the Bronze Age eruption of Thera in the Aegean Sea.



Plate 2.5 Pollen grains of *Chenopodium album*, an indicator of open-ground conditions, ~40 microns (0.04 mm) in diameter.



Most pollen analyses are carried out at sites – such as lakes and bogs – which have experienced continuous sediment accumulation, although source materials as unusual as Ice Age hyaena dung have proved to contain pollen (Carrión et al., 2007)! A series of pollen samples from different depths in a sediment profile are normally presented as a pollen diagram, which conventionally shows depth or age on the vertical axis and frequency on the horizontal axes and which can be drawn up using specialist graphics software linked to programs such as TILIA and C2. Pollen and other microfossil diagrams are used extensively in this book, in some cases based on absolute counts, in others based on proportional ones. The diagrams sometimes show only selected taxa, or major pollen groups such as arboreal (tree) versus non-arboreal pollen (AP vs. NAP). [Figure 2.10A](#) is a percentage pollen diagram from a site in southern France, which shows changes in the main tree types against core depth. As the calibrated ^{14}C dates indicate, this section of the core covers a time period from the Late Glacial through to the middle of the Holocene. Like most pollen diagrams, it is subdivided into zones within which samples (or spectra) are similar in their essential characteristics. Before the advent of radiocarbon dating, these zones were assumed to be synchronous over whole regions. In the British Isles, for instance, Sir Harry Godwin proposed a pollen zonation scheme that began in the Late Glacial (zones I–III) and continued through the Holocene to the present day (IV–VIII). Regional pollen zones are now known to be time-transgressive and are therefore used less widely than previously, but local pollen assemblage zones remain an essential part of individual site investigations. These local pollen zones are often designated with the help of statistical methods such as cluster analysis (Birks et al., 2012).

Figure 2.8 Vegetation reconstruction based on pollen analysis (see text for explanation).

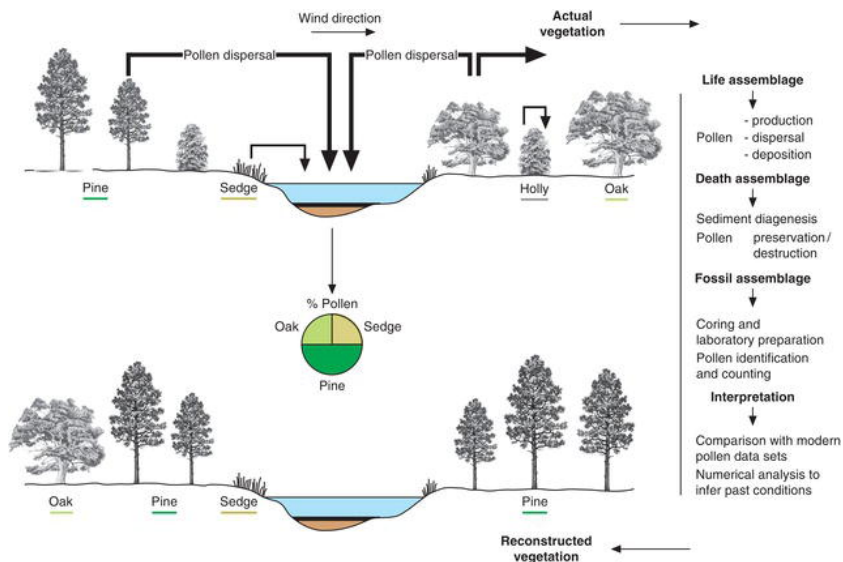


Table 2.5 Absolute and relative pollen counting statistics

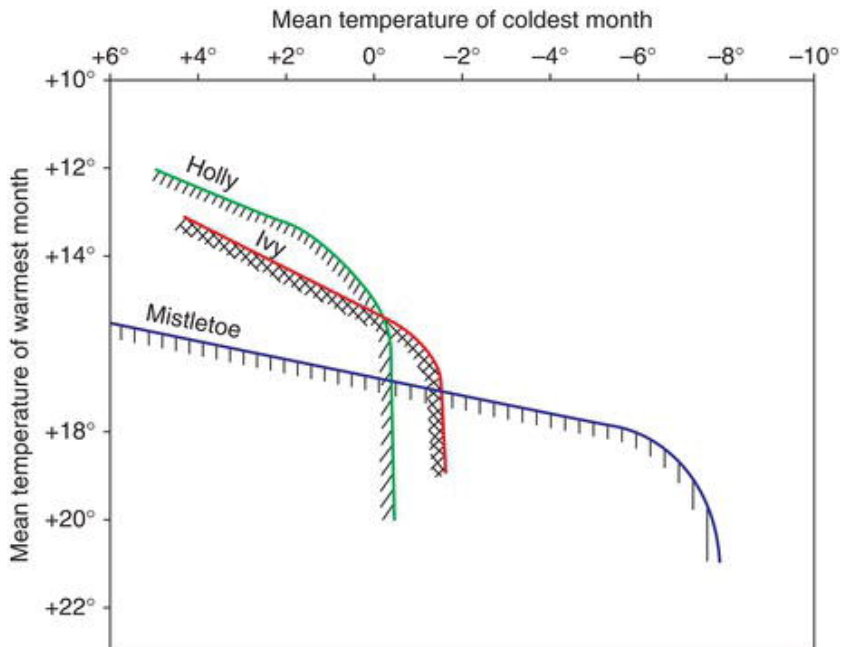
	Absolute numbers			Relative proportions		
	Pine	Oak	Sedge	Pine (%)	Oak (%)	Sedge (%)
Time 1	10	5	5	50	25	25
Time 2	10	5	35	20	10	70

The initial objective of a pollen analytical study is normally to reconstruct past vegetation and how it has changed over time (Mitchell, 2011), but there is often a second objective – that of establishing the factors which determined the former flora. Two of the most important external controlling factors are **climate** and **human activities**. Assuming little human disturbance, it is widely believed that vegetation will eventually achieve a state of equilibrium with the prevailing climate (Webb, 1986; Edwards and MacDonald, 1991a; Ritchie, 1995). If this assumption is correct, then modern data on climate–plant relationships can be applied to pollen-based vegetation histories to produce estimates of past temperatures or rainfall levels (Birks and Birks, 2003). Take, for example, the arctic/alpine plants dwarf birch (*Betula nana*) and mountain avens (*Dryas octopetala*), whose pollen and macrofossils are common in northern European Late Glacial assemblages. In the British Isles, dwarf birch is today virtually restricted to the Scottish Highlands, where mean summer temperatures are below 22°C (Connolly and Dahl, 1970). But before accepting this as the temperature of a typical July day in southern England 12 000 years ago, it should be remembered that species distribution is a function of more than one climatic variable. In an attempt to overcome this problem, Johannes Iversen (1944) combined two

climatic variables and several different indicator species. He showed, for example, that holly can tolerate cool summers but not cold winters, whereas mistletoe (*Viscum album*) is the reverse, requiring summer warmth but being relatively hardy in winter. Ivy (*Hedera helix*) lies somewhere between the other two in its climatic tolerances (see [Figure 2.9](#)). Using the overlapping ranges of the three shrubs in combination, it is possible to provide more precise temperature reconstructions than if each were used on its own and to provide an indication of past climatic continentality.

To extend this approach to a ‘multivariate’ analysis of vegetation and climate requires the use of more complex statistics, such as transfer functions. These use a ‘training set’ of sites spanning a range of different environments, and which relates modern species assemblages to the main climatic controls. Numerical multivariate techniques were initially developed to reconstruct past climate from pollen data by Tom Webb (1985), Pat Bartlein and colleagues (Bartlein et al., 1986). This collapsed the normal two-step reconstruction of pollen to vegetation to climate into a one-step process, involving the creation of pollen–climate ‘response surfaces’. Climatic calibration of modern and fossil pollen data has allowed maps to be produced showing temperature, precipitation or air mass distribution for different times in the past over eastern North America (Webb et al., 1993) and other regions of the world (e.g. Bonnefille et al., 1990, for Africa). However, in regions such as Europe, where human impact on vegetation has a longer antiquity, pollen-based reconstruction of climate is more uncertain, at least for recent millennia. Another difficulty arises when there is no good modern analogue for a mixture of species that coexisted at some point in the past. There are risks of over-interpreting data using the transfer function approach, especially if trying to infer more than one variable from the same data set at the same time, for example, precipitation as well as temperature (Huntley, 2012; Juggins, 2013). The ‘space for time’ (or ergodic) approach to calibration has been successfully applied to aquatic organisms as well as to pollen (Birks, 1998; Birks et al., 2012; see following text), but it does not work for all types of data, for example, when linking tree-ring widths to prevailing weather conditions (see [Technical Box X](#)).

Figure 2.9 The thermal limits of holly, ivy and mistletoe (after Iversen, 1944). Reproduced with permission of the Editor of GFF and the Geological Society of Sweden.



The interpretation of pollen records in terms of past human impact is, if anything, more difficult than it is for climate (Edwards and MacDonald, 1991b). Many former land uses, such as hunting–fishing–foraging in Mesolithic Europe, have no modern analogue for comparison. Furthermore, it is often the indirect consequences of human impact that are most obvious palynologically (Edwards, 1979, 1982). For example, the pollen of wheat, barley and other cereal crops is poorly dispersed and only weakly represented in most European pollen diagrams until historical times. More prominent evidence of prehistoric agriculture instead comes the pollen of ruderals (weedy plants) like docks (*Rumex* sp.) and nettles (*Urtica* sp.) (Turner, 1964; Behre, 1981, 1986). These species take advantage of disturbed ground regardless of cause and are therefore not diagnostic indicators of anthropogenic disturbance. Many, for example, were relatively abundant during the Late Glacial before a stable forest cover had been established (Godwin, 1975; Huntley and Birks, 1983, p. 518ff.).

The recognition of human disturbance in pollen diagrams usually involves a combination of evidence, including changes in the overall composition of vegetation as well as indicator species (Maguire, 1983; Birks et al., 1988). Among the latter would be grassland indicators such as ribwort plantain (*Plantago lanceolata*), while the former would include changes in the AP : NAP ratio. Pollen theory (e.g. Prentice, 1985) predicts that small sites like woodland hollows recruit most of their pollen from vegetation growing locally. In contrast, large sites, such as lakes >1 km² in area, receive most of their pollen from regional rather than local sources. Based on this, Shinya Sugita (2007a,b) developed computer models – rather exotically called LOVE and REVEALS – for calculating the contributions made by different plants to the pollen deposited at sites of different size, so long

as the pollen productivity of each plant type is known (cf. [Table 2.4](#)). As a result, it is now possible to estimate quantitatively the proportion of individual plant types living in a defined area around a pollen core site at different times in the past (Nielsen and Odgaard, 2005; Gaillard et al., 2010). When coupled to a local GIS model, these estimated plant abundances can be placed back where they are most likely to have grown on the local or regional landscape (Fyfe, 2006). In other words, these approaches can take us much closer to reconstructing real landscapes of the past, showing which areas were forested, farmed or under other land uses, even for prehistoric times (Fyfe et al., 2010). However, in some cases, a former vegetation change might be attributed to more than one cause: the mid-Holocene European elm decline, for example, has been variously attributed to prehistoric agriculture, disease, climate change or a combination of more than one of these factors; this is an example of [equifinality](#). It is one of the tasks of palaeo-science to test between competing causal hypotheses.

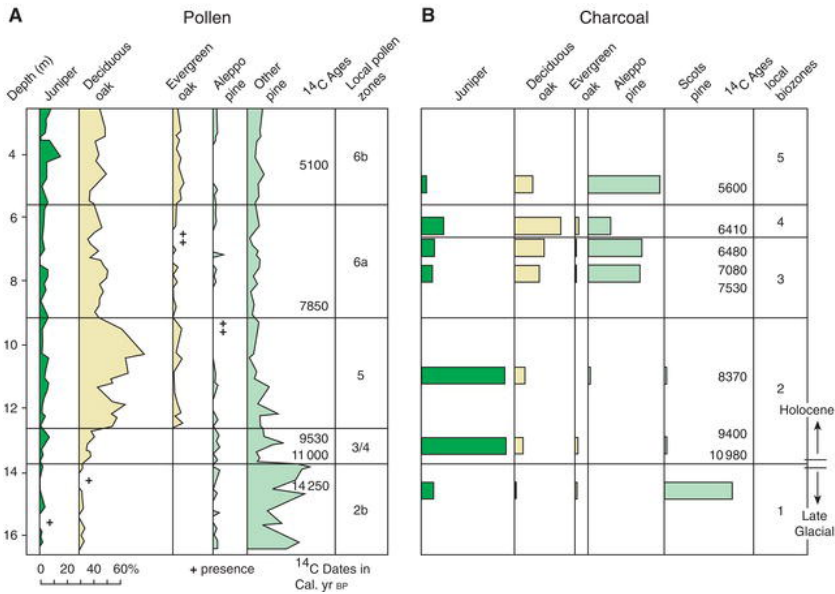
Plant remains

Notwithstanding the enormous success of [palynology](#) in reconstructing former vegetation, climate and land use, pollen research alone cannot hope to answer all our questions about past plant ecology. The value of alternative, complementary sources of information is well illustrated by the example of plant macrofossils, that is, plant remains like seeds, leaves and wood that are visible with the naked eye. Most of the same principles examined in relation to pollen – life and death assemblages, and so forth – apply also to plant macrofossils, and indeed to other organisms. However, bigger sample volumes are required because macrofossils are larger than pollen, and counting statistics are more difficult because of the wider variety of parts of the plant that are preserved. Compared with pollen, however, plant macrofossils have a number of advantages (Birks and Birks, 2003). They form part of the once living plant rather than the reproductive stage of the life cycle. As a result, they are usually identifiable to species level and the ecological inferences that can be drawn from them are relatively precise. Unlike pollen grains, macrofossils are not transported far from their point of origin, and their local presence can be assumed with greater confidence. They can also help record plants with poor pollen production or dispersal, such as the arctic flower mountain avens (*D. octopetala*).

On archaeological sites, macrofossils are often found in the form of charred seeds or wood charcoal which offers a direct indication of the plant resources exploited by former human communities (Pearsall, 2000). The study of wood charcoals – or anthracology – can offer useful insights into localized vegetation changes which complement those provided by pollen analysis (e.g. Emery-Barbier and Thiébault, 2005). [Figure 2.10](#) compares the Late Glacial to mid-Holocene woodland history from a charcoal record in an excavated prehistoric rockshelter in southern France, with that of a pollen core from a valley mire nearby. The latter represents a more continuous record of regional vegetation change, but there are problems with interpretation of some of the pollen data which the charcoal record can help to resolve. For instance, the Late Glacial period before c.11 500 Cal. yr BP has high pine pollen percentages at Tourves, but does this

really mean that pine trees were common in the vicinity? Pine is notoriously well-dispersed by the wind and often makes up a significant proportion of the total pollen found in open, exposed landscapes such as tundra, even though it may not be growing locally (Ritchie, 1987). Fortunately we know in this case that pine trees really were present – and Scots pine at that – because the Epipalaeolithic occupants of the rockshelter collected it at the time (Figure 2.10B). On the other hand, changes in the charcoal record could also have been caused by cultural preferences, for instance, because one type of wood burns better than another. The switch from juniper to oak and pine charcoal just before 7500 Cal. yr BP, for example, coincides with the first appearance of Neolithic farmers in this area. This apparent vegetation change may reflect the fact that oak and pine woods were being cut down to make way for cultivated crops, rather than because juniper – which had been used previously – became less common in the regional vegetation. Charcoal particles are also found in lake sediments and peat bogs (Patterson et al., 1987; Whitlock and Larsen, 2001; Conedera et al., 2009), but as microscopic, rather than macroscopic, fragments. While they are normally too small to permit identification to tree taxa, these tiny charcoals do offer very important information about fire histories (e.g. Bennett et al., 1990; Colombaroli et al., 2008; Turner et al., 2008).

Figure 2.10 Comparative records of Late-glacial and early Holocene changes in selected trees from southern France; A pollen record from a valley peat bog at Tourves (modified from Nicol-Pichard, 1987; reproduced with permission of the University of Avignon); B wood charcoal record from the nearby prehistoric rockshelter site of Fontbrégoua (data from Thiébault, 1997).



Plant macrofossil analysis has been applied to a range of palaeoecological problems. In western Norway, Hilary Birks (1993) found some striking differences between the vegetation records indicated by pollen and plant macrofossils in the same lake sediment cores. In particular, birch (*Betula*) was prominent in the

et al., 1990; Scott, 1990).

Many plants form microscopic **phytoliths** (literally ‘plant stones’), usually made out of silica. Being robust, they survive well in archaeological deposits, soils and lake sediments, and they can be used to reconstruct past vegetation and botanical resources even though the parts of the plant made out of carbon have been oxidized and lost (Piperno, 2006). This technique has been especially useful in tropical and dryland regions, where high rates of evapotranspiration encourage phytolith formation, particularly in monocotyledon plants such as reeds and cereal grasses. Soil and water conditions during the growth period will also affect the number and types of plant cells that become silicified. This has permitted researchers such as Arlene Rosen to distinguish between cereal crops cultivated under irrigation and those grown through rain-fed cultivation, right back to Neolithic times (Rosen and Weiner, 1994).

Creatures great and small

In some situations, proxy data based on fauna rather than flora can prove most informative, and the potential range of animal indicators is very wide – from mites to mammoths! At the smaller end of the size scale, land snails (**mollusca**) have proved of great value in the investigation of land-use and vegetation history, especially on calcareous rocks where pollen sites are almost absent, such as the English chalk. Molluscs can be extracted without chemical pretreatment, and identification to species level is generally possible. The species composition of land snail assemblages is largely dependent on climate and on local habitat, especially vegetation cover. During the terminal Pleistocene and early Holocene, land snails found in colluvial and similar sequences closely mirror vegetation changes over the same time period (Kerney, 1977; Preece, 1993). Later in the Holocene, land snails have proved especially valuable in archaeological contexts, such as soils buried beneath burial mounds (Evans, 1972, 1993; Thomas, 1985; Davies, 2008). Species which are shade tolerant or intolerant provide good indicators of whether the vegetation was formerly wooded or open and hence of forest clearance and regeneration.

Beetles (**Coleoptera**) are another important type of organism which can respond rapidly and sensitively to changes in climate. Among the beetles, there are about 350 000 different species – more than all the flowering plants combined – so, not surprisingly, many are adapted to highly specific ecological niches. For example, many species have fastidious preferences for warmth or cold. Taxonomy is based on the characteristics of their external (or exo) skeletons which are often well preserved in sediments, but which become disaggregated into their component parts (head, thorax, etc.) (see Plate 3.1). Coleoptera have helped, in particular, in the reconstruction of late Pleistocene climates in mid- to high-latitude regions such as northwest Europe (Coope, 1970; Elias, 2001). Their thermal likes and dislikes, coupled with an ability to migrate rapidly, make beetles ideal indicators of past temperatures. Beetle assemblages have been climatically calibrated for both summer and winter temperatures using the mutual climatic range method (Atkinson et al., 1987), similar to the method developed

by Iversen for plants (see [Figure 2.9](#)). Holocene climatic reconstructions based on Coleoptera have so far proved more problematic, although beetle remains are valuable in indicating vegetation openness and grazing intensity by animals (Whitehouse, 2006).

Insects which have to spend the early part of their life cycle living in water such as Trichoptera (caddisflies) and chironomids (midge larvae) are other valuable sources of information about past environmental conditions (Williams, 1988; Walker et al., 1991b; Elias, 2001; Walker, 2001). Chironomids are just as sensitive to warmth and cold as beetles, and their preserved head capsules have been used to reconstruct Late Glacial temperature fluctuations in regions such as eastern Canada and northwest Europe (Walker et al., 1991a; Brooks et al., 1997, 2007; Heikkilä and Seppä, 2003). They have also provided an index of changing lake water quality later in the Holocene (Sadler and Jones, 1997; Langdon et al., 2006). A different type of insects, namely, mites, can provide an indirect proxy for the abundance of grazing animals in a landscape (Chepstow-Lusty et al., 2007).

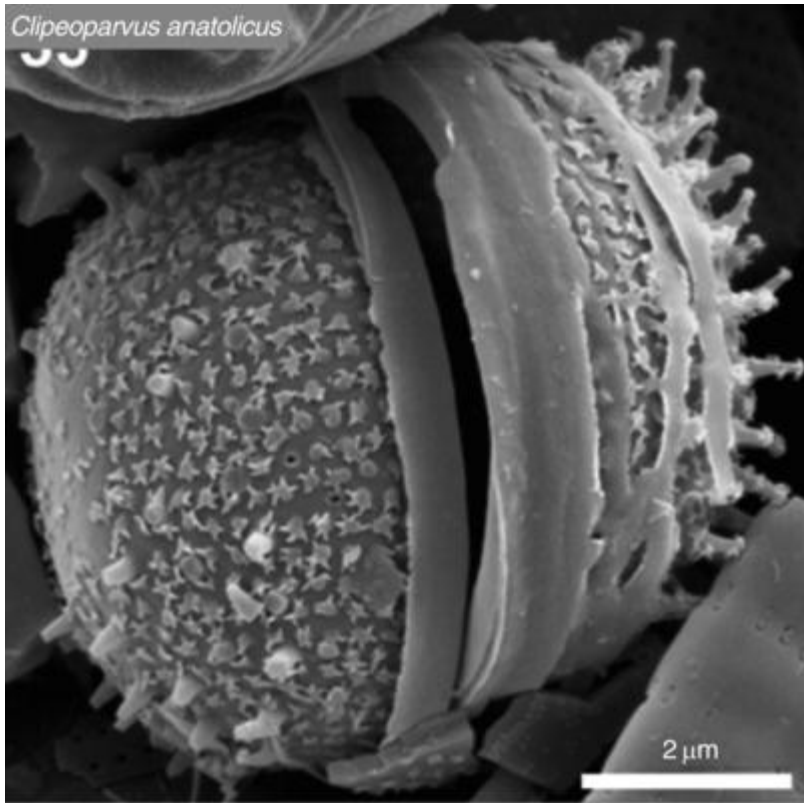
Faunal analysis is not restricted to tiny creatures, however, but also includes much larger ones, notably mammals, reptiles, fish and birds. With complex organisms such as these, there are issues about how to interpret individual fossil counts, normally requiring the investigator to calculate the minimum number of individual creatures that were present. For example, a left front tibia and a right front tibia from a mixed deposit could have come from the same animal or from two different ones, whereas two left front tibia bones would provide proof that two individuals had existed. The study of animal bones on archaeological sites – or archaeozoology – provides one of the most important lines of evidence about past human economies, including livestock domestication (see Chapter 5). In non-archaeological sites such as the Rancho La Brea Tar Pits in Los Angeles, the bones of sabre-toothed big cats, wolves, mammoths and other megafauna provide our main evidence about the great beasts that became extinct at the end of the Last Ice Age (see Chapter 3). One rather special type of large mammal is *Homo sapiens*, the only species of ‘human’ to have been living on the planet during the last 20 000 years. Human skeletal remains are most commonly found as deliberate burials at, or near, archaeological sites. As well as anatomical study, human bones – and indeed all biological remains – can potentially be analysed chemically and for their ancient DNA. Alongside ancient DNA, the genetic make-up of modern biological populations also gives us clues about their past history. For human remains, the Y-chromosome provides a history of every individual’s paternal line (father, grandfather, etc.), while mitochondrial DNA gives information about maternal origins (Oppenheimer, 2004; Brown and Brown, 2011). For plants, biomolecular evidence can tell us about the likely place of origin of domestic crops such as wheat and barley (Heun et al., 1997; Jones and Brown, 2000) or about how tree species population dynamics have responded to Quaternary environmental change (Hofreiter et al., 2012).

Freshwater and marine organisms

Past changes in the oceans, the coastal zone and freshwaters like rivers and lakes can be gleaned from the remains of creatures which once lived there (Smol et al., 2001; Haslett, 2002). They range from marine zooplankton such as Foraminifera to non-marine Protozoa such as testate amoebae (Charman et al., 2000), through ostracods (Holmes and Engstrom, 2003) to the pigments left by phytoplankton (Leavitt, 1993). Some of these organisms make their external shell out of calcium carbonate, as molluscs do. Bivalve marine molluscs like mussels and oysters are an important source of free food along the shoreline which have been exploited for thousands of years, as testified by ancient shell middens (essentially large piles of discarded shells) that are found along coasts as diverse as Peru and the Scottish islands. Not surprisingly, fish bones and otoliths (ear bones) are also commonly found in coastal archaeological sites.

In **palaeolimnological** studies, aquatic indicators such as **diatoms**, chrysophytes and cladocera (water fleas) have proved especially valuable (Battarbee, 1986; Korhola and Rautio, 2001; Smol and Stoermer, 2010). Diatoms (Bacillariophyceae) are unicellular algae with a shell (or frustule) made of silica. The frustule comprises two, usually symmetrical valves, which are finely sculptured and ornamented with rows of tiny holes (see [Plate 2.6](#)). Diatom taxonomy is based on the shape, size and ornamentation of the frustule. Although there are several other types of algae, diatoms are one of the few to be preserved in any detail after death. The frustule of all except a few fragile species survives well in sediments and allows taxonomic identification to species or even subspecies level. Diatoms are about the same size as pollen grains, and in many ways diatom analysis fulfils the same function for aquatic ecosystems that palynology does for terrestrial vegetation. Compared to pollen, diatoms have two advantages: firstly, the higher taxonomic resolution possible and secondly, the fact that diatom fossils represent the organism itself rather than a part of the reproductive process, with all the biases that this avoids. Diatoms occupy an extremely diverse set of habitats, including not only freshwater lakes and rivers but also saline waters such as closed lakes, estuaries and the ocean itself, and they are even found in peat and soil. Using modern species and community ecology, it is possible from fossil diatoms assemblages to reconstruct past pH, salinity, water depth, productivity and other conditions via statistical transfer functions (Birks et al., 1990, 2012; Fritz et al., 1991; Reid et al., 1995; Moser et al., 1996; Gasse et al., 1997; Smol and Stoermer, 2010).

Plate 2.6 Diatom frustule of the small centric species *Clipeoparvus anatolicus*. © Natural History Museum, London. Reproduced with permission.



Geological techniques

Recent earth history can be understood by examining sediment sequences and landforms at the Earth's surface. Those from terrestrial environments (i.e. on land) can be classified into a series of different geomorphological regimes, each with its own characteristic set of depositional environments (see [Table 2.6](#)). As in palaeoecology, the reconstruction of past geological environments is based on uniformitarian principles; in other words, our knowledge of the present is used as the key to understanding the past. Modern data about earth surface processes are applied to fossil landforms and sediments, firstly to assign them to a particular geomorphological regime and secondly to interpret their origin in terms of past processes and climates. No analogue situations are encountered less often than in palaeoecology, although they do sometimes occur, notably for high-magnitude events. The so-called Channelled Scabland of Washington State, for instance, was for many years explained in terms of observable modern fluvial processes (Baker and Bunker, 1985). It is now realised that it was in fact created by catastrophic, short-lived flooding when ice-dammed Lake Missoula burst at the end of the last

glaciation.

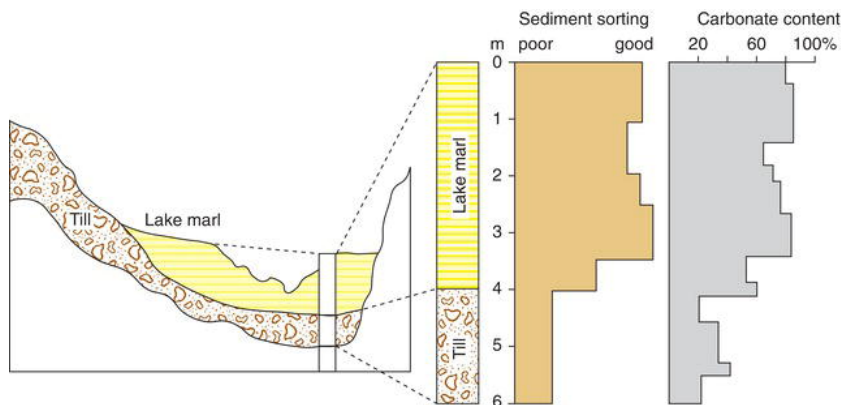
Table 2.6 Characteristics and examples of different geomorphic regimes

	Sedimentation	Sediments	Landforms
Lacustrine	Continuous (lake bed)	Lake marl	Shoreline terrace
Fluvial	Semi-continuous	River gravel	Palaeo-channel
Coastal marine	Semi-continuous	Beach sand	Raised beach
Aeolian	Semi-continuous	Dune sand	Fossil dune
Glacial	Discontinuous	Till	End moraine
Mass movement	Discontinuous	'Head'	Solifluction lobe

Landforms can be produced by erosion, deposition or a combination of both types of process. Terrace features, in particular, are formed by a phase of sedimentation – say, by an aggrading river – followed by a period of erosional downcutting. Terraces provide valuable information on previous river, lake and sea levels that can be recorded by survey and mapping. However, it is possible to misinterpret the origin of terraces and other landforms unless data on surface morphology are supplemented by an analysis of the underlying sediments. Earth histories can therefore be studied most effectively by combining landform-based techniques with those based on sediment stratigraphies (Goudie, 1990), as illustrated by the sequence shown in [Figure 2.12](#). In this case, a glacial moraine is overlain by lake sediments, now being dissected by fluvial erosion. Of these three geomorphic regimes, only the first two have left sedimentary records, and their contrasting physical and chemical characteristics are recorded on the right-hand side of the diagram.

Minerogenic or clastic sediments, such as river gravels, comprise rock particles of allochthonous origin – that is, they were brought to the point of deposition from elsewhere. Analysis of their physical characteristics involves examining particle shape, size distribution and arrangement in space. One standard measurement technique is particle size analysis, from which size and sorting indices may be derived (Folk, 1980). The characteristics of two contrasting clastic sediment types are listed in [Table 2.7](#). Till and dune sand are easily distinguished from each other, but differentiation may be harder in other cases. Soliflucted ‘head’ has similar properties to some glacial tills, for instance, while desert dune and beach sand could be confused on the basis of sediment characteristics alone. Minerogenic deposits can be analysed for their mineral magnetic properties, for their clay and ‘heavy’ mineral compositions and for the concentration of individual chemical elements, such as iron and titanium. The latter are measured by techniques such as X-Ray Diffraction and X-ray fluorescence (XRF), including rapid, non-destructive XRF methods like ITRAX scanning (Croudace et al., 2006).

Figure 2.12 Environmental reconstruction based on landform and sedimentary evidence.



Precipitates, such as lake marl, are sediments chemically deposited from aqueous solution and are therefore **autochthonous** in origin. Most commonly, deposition takes place in standing water conditions as is encountered in lakes and the sea, and it involves one of two main mechanisms. The first of these is biochemical ‘fixing’ of solutes by aquatic or marine organisms, like the polyps responsible for building up coral reefs. The second is direct chemical precipitation from water, especially under saline-alkaline conditions where solutes are highly concentrated, for example, in a playa lake or coastal lagoon (see [Plate 2.7](#)). The type of salt that is deposited depends on the chemical composition and concentration of the water (Eugster and Kelts, 1983). Another type of chemical precipitate is speleothem carbonates found in caves, including stalagmites and flowstones. Like other precipitates, speleothems can be analysed isotopically (see following text).

Table 2.7 Comparative properties of glacial and aeolian sediments

	Glacial till	Dune sand
Particle shape	Angular	Spherical
Grain surface	Fractured	Smooth
Size distribution	Mix of fine and coarse particles	Homogenous sand
Particle sorting	Very poor	Good
Sediment structures	Weak or absent	Obvious cross bedding
Particle alignment	May be parallel to direction of ice flow	Not evident

Plate 2.7 Chemical sediments precipitated from a hyper-saline lake on the Anatolian plateau.



Biogenic sediments are comprised primarily of organic carbon compounds and are usually autochthonous, for example, in a raised peat bog. Less commonly, organic matter may be allochthonous, as when soils are eroded and washed into streams and lakes. Organic matter content can be measured using methods such as loss on ignition, which provides a useful measure of the amount of carbon in soils and sediments. The ratio between carbon and nitrogen provides an index of the source of organic matter, because soils, algae and higher plants all have different C : N ratio signatures. Organic matter is, of course, important for radiocarbon dating and also for stable carbon isotope analysis (see following text). Biogenic sediments typically contain abundant micro- and macrofossils and are therefore extensively used in pollen and other palaeoecological investigations (Chambers et al., 2012). In addition, the organic component of sediments can be analysed chemically for specific marker compounds (Castaneda and Schouten, 2011) or by methods such as near-infrared spectrometry (NIRS) which is quick and non-destructive (Rosén, et al., 2000).

In practice, most sediments comprise a mixture of clastic, chemical and biogenic elements. One way in which the differing proportions of elements making up a sediment sample can be assessed is by using a system of sediment classification, such as that developed by the Danish palaeoecologist J. Troels-Smith (Birks and Birks, 1980, p. 38ff.; Aaby and Berglund, 1986). Of Troels-Smith's six main sediment categories, three are biogenic, a further two are minerogenic and a further one is either chemical or biogenic (see [Table 2.8](#)). Although this semi-quantitative system of sediment classification is widely applied, its usage is restricted to freshwater lake and mire sediments in humid temperate regions. Lakes and mires provide some of the most accessible and best-studied sediment 'archives' recording Holocene environmental change, with the former being especially informative when the sediments are annually laminated, or varved (Zolitschka, 2003; Ojala et al., 2012). Sediment in small temperate-zone lakes and mires typically builds up at about 1 m/1000 years, so that a full

Holocene record is most likely to be represented by around 10 m of accumulated sediment (Webb and Webb, 1988).

Table 2.8 Elements of the Troels-Smith sediment classification.

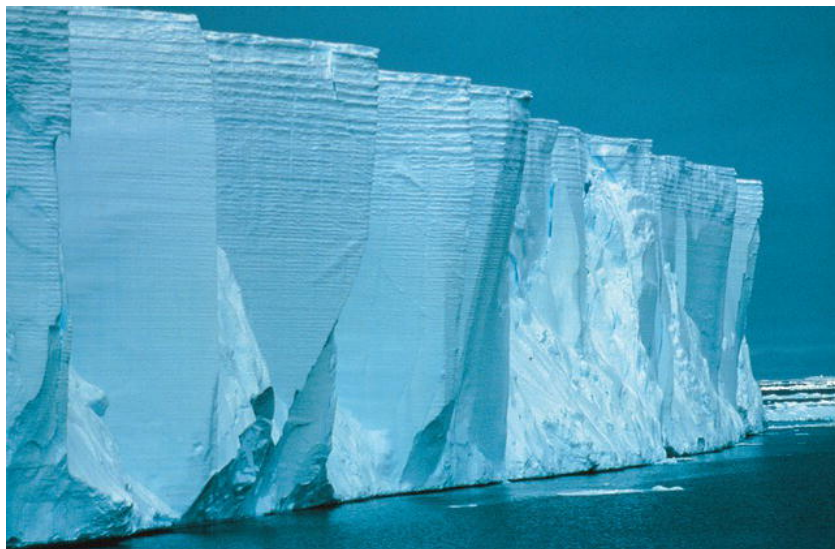
- a) *Substantia humosa*: undifferentiated disintegrated organic matter
- b) *Turfa*: macroscopic plant remains of below-ground origin and mosses, for example, roots
- c) *Detritus*: macroscopic plant remains of above-ground origin, for example, leaves
- d) *Limus*: lake mud of biogenic or chemical origin, for example, marl
- e) *Argilla*: fine-grained minerogenic particles, for example, clay
- f) *Grana*: medium- or coarse-grained minerogenic particles, for example, sand

Ice and ocean

The deep ocean bed is a depositional environment which has proved invaluable for understanding global change during the Pleistocene ice age. Deep-sea sediment cores contain a record of long-term fluctuations between cold glacial and warmer interglacial climates (Lowe and Walker, 1997). The generally slow rate of sedimentation is ideal for studying such long spans of time, although it means that the uppermost sediments formed during the Holocene are often no more than a metre thick. This does not allow the kind of detailed analysis of change through time that is possible in lake sequences, although in enclosed seas like the Mediterranean and in other marine basins, sediment accumulation rates have often been faster than in most of the open ocean. Here, pollen and other analyses can help link deep-sea core records to those on land (Rossignol-Strick, 1995; Sánchez-Goñi et al., 2002). In a few cases, deepwater **anoxia** has allowed the formation of varved sediments, as it has in some lakes; a good example is the Cariaco Basin off the coast of Venezuela. Here, a continuously varved marine sediment record provides an unbroken high-resolution history of climatic and ocean conditions throughout the Holocene and back into glacial times (Hughen et al., 1996). Changes in sea-surface temperatures can be reconstructed using the composition of foraminifera and other fossil organisms found in marine cores, or from biochemical indicators. For example, TEX86 (tetraethers consisting of 86 carbon atoms) provides a proxy for past temperatures in the sea and lakes (Schouten et al., 2002; Powers et al., 2004), as does the alkenone UK'37 (Rostek et al., 1993; Emeis et al., 2000; Eglinton and Eglinton, 2008). One unique organic

compound produced by algae, labelled IP₂₅, provides a proxy for past sea ice extent in polar regions (Belt et al., 2007).

Plate 2.8 Annual snow accumulation layers, thicker at the surface and compressed with depth, are clearly visible in the edge of the Antarctic ice. These annual bands provide a ready-made timescale for ice core records of past climatic change. Reproduced with permission of Charles Swithinbank.



At the opposite extreme to the deep ocean, the Holocene stratigraphic record preserved in ice sheets is not metres but thousands of metres thick! Cores drilled through the layered accumulations of snow and ice deposited in the polar or high-mountain ice sheets permit an extraordinarily detailed reconstruction of past climate (Plate 2.8). The longest records, from sites such as Vostok, Dome C and Dome Fuji in Antarctica, extend backup to eight full glacial–interglacial cycles (Jouzel et al., 1993; EPICA Community Members, 2004). The ratios of hydrogen and oxygen isotopes in the ice provide an index of former temperatures, while the dust content indicates global wind strength and aridity, and its acidity provides an index of major volcanic eruptions (Zielinski et al., 1994). Furthermore, snow, as it accumulates and is compressed into ice, traps air bubbles whose methane (CH₄) and CO₂ contents can be measured from the cores to provide a record of the changing concentrations of these greenhouse gases (Lorius et al., 1990). This applies not only to long timescales when the climate varied due to natural factors but also to changes in recent centuries associated with fossil fuel burning and other human impacts.

Ice does not only build up vertically but also flows out laterally from the centre of the ice sheet to its margins. For this reason, cores taken at the top of the ice sheet dome should have experienced less deformation and contain a longer record than those taken near the edges. On the other hand, the high points of ice sheets are among the least accessible and most inhospitable places on Earth; Camp Vostok in Antarctica has a mean annual temperature of -55°C and stands 3490 m above sea level. In the case of the Huarascán ice cap on the top of the

Peruvian Andes, Lonnie Thompson and his team had to manually haul six tonnes of solar-powered drilling equipment up to over 6000 m altitude (Thompson et al., 1995)! After he had cored the small, shrinking ice cap at the top of Kilimanjaro in Africa, he brought his ice cores down by hot-air balloon....

In the northern hemisphere, Greenland contains the most important ice core record. Ice has accumulated faster here than in Antarctica, and it therefore offers a more detailed history of climatic change. In a competition reminiscent of Amundson and Scott's early twentieth-century race to the South Pole, two separate teams – one European, the other American – drilled cores only 28 km apart at the summit of the Greenland ice sheet between 1988 and 1993. (The Europeans 'won' when they reached bedrock at 3028 m on 12 July 1992, just under a year ahead of the Americans.) The duplication of effort in producing two ice cores, called GRIP and GISP2 respectively, had important benefits, because it allowed a cross-check to be made; this fortunately revealed excellent agreement, except for the last interglacial (Taylor et al., 1993). In the Greenland Summit cores, the Holocene is represented by 1.6 km vertical thickness of ice, permitting annual time resolution and providing a very important record of northern hemisphere climate.

Stable isotope analysis

As noted earlier in this chapter, most chemical elements have a few rare forms with the 'wrong' number of neutrons. Some of them, like ^{14}C , are isotopically unstable and undergo radioactive decay with time, but others are stable and provide information instead on past environmental conditions or as environmental tracers. Among the most important are the isotopes of carbon and oxygen, while others that have been used include those of nitrogen, sulphur and strontium.

The ratio between the heavy isotope oxygen-18 (^{18}O) and the lighter, and much more common, oxygen-16 (^{16}O) varies with temperature, and this fact led its discoverer – Harold Urey – to claim he had a 'geological thermometer' in his hands. Cesare Emiliani measured this $^{18}\text{O}/^{16}\text{O}$ (or $\delta^{18}\text{O}$) ratio in the fossil shells of tiny marine planktonic foraminifera to try to reconstruct changes in Pleistocene temperatures. The oxygen isotope stages that he devised now form the way that the Quaternary Ice Age is classified into episodes of warmer and colder climate (see [Technical Box IV](#)). Subsequently, Nick Shackleton and others established that a more important direct factor than temperature in influencing the marine isotopic record was the growth and decay of ice sheets. These lock up isotopically 'light' water as ice and leave the oceans enriched in the heavier isotope ^{18}O . In other words, oxygen isotope analysis of deep-sea cores provides an index mainly of past glaciation rather than of temperature. Temperature has been a more important control over the $\delta^{18}\text{O}$ record in some other systems, such as freshwater lake sediments, ice sheets and speleothems. Oxygen isotopes can indicate seasonal as well as annual variations in temperature or rainfall, for example, in mollusc shells as they accrete (Abell et al., 1995). In lakes from dryland regions, the stable oxygen isotope ratio varies primarily according to the intensity of evaporation

from the water surface, where they reflect past changes in hydrological balance (Talbot, 1990; Leng and Marshall, 2004). Most oxygen isotope measurements are carried out on carbonates, either biogenic, like ostracod shells, or authigenic, such as speleothems. In acidic environments, carbonates are usually not preserved, but here there has been progress in using chironomids and biogenic silica from diatoms and phytoliths for $\delta^{18}\text{O}$ measurement (Wooller et al., 2004; Leng and Barker, 2006).

Stable carbon isotope analysis provides a way to determine past changes in the type of plants and changes in the global carbon cycle, for example, as the atmospheric concentration of greenhouse gases has fluctuated. Some plants, such as tropical grasses and sedges, incorporate CO_2 as a four-carbon molecule during photosynthesis, the so-called C4 pathway. Other grasses, along with trees and shrubs, use a three-carbon (C3) pathway. The relative importance of C4 plants can be established by measuring the ratio of two stable isotopes of carbon, ^{13}C and ^{12}C . That ratio, expressed as $\delta^{13}\text{C}$, has been analysed in plant macrofossils, soils, ostracods, land snails, timbers and peats (Heaton et al., 1995; Street-Perrott et al., 1997; Douglas et al., 2012). In lake sediments, the stable carbon isotope ratio is affected by the productivity of aquatic algae, as well as the type of vegetation in the lake catchment (Lane et al., 2004), while in arid environments, $\delta^{13}\text{C}$ analysis of land snail shells has been used to reconstruct past changes in rainfall (Goodfriend, 1992). Carbon and nitrogen isotopes have also been used to help determine past changes in human (and animal) diet. Maize, along with some other grain crops, incorporates CO_2 via the C4 pathway (Chisholm, 1989; Richards et al., 2003), and the relative dietary importance of C4 plants can be established by measuring the $\delta^{13}\text{C}$ and $\delta^{15}\text{N}$ ratios in bones of humans or livestock. This ratio is high in the case of diets with a large proportion of C4 plants, such as maize, or of animals fed off C4 plants. The $\delta^{13}\text{C}$ and $\delta^{15}\text{N}$ values are low when C3 plants, including wild foodstuffs, dominate the food chain.

Geomorphology and climate

Climatic conditions exert a powerful influence over geomorphic processes and landforms. Geomorphologists such as Julius Büdel (1982) argued that at a global-scale distinct morphoclimatic regions can be recognised, within which a common set of processes are dominant. Thus ice-wedge polygons are exclusively found in the periglacial zone of winter freezing and summer thaw, where the average annual temperature is below -6°C (Harris and Murton, 2005). Geomorphic processes such as freeze-thaw are so closely linked to climate that it is possible to calibrate them against temperature or precipitation. However, not all landforms are harmoniously adjusted to the climate prevailing today because some were produced in the past when climatic conditions were different. These relict landforms may furnish indications about past or palaeoclimates.

One geomorphic regime intimately associated with specific climatic conditions is the **glacial** environment. Valley glaciers form under cold, moist climates, specifically in areas where the accumulation of winter snow exceeds summer melting or ablation. The surplus snow builds up, becomes compressed to

form ice and moves downslope by basal sliding or flowage. In due course, the glacier enters a zone of warmer air temperatures in which there is a net deficit between accumulation and ablation, and eventually the glacier terminates. The mass balance of a land-based glacier is therefore a function of winter snowfall on the one hand and summer temperature and cloudiness on the other (Evans, 2003). Under colder or more snowy conditions, the glacier snout will advance down-valley; under warmer or drier conditions, it will retreat. On the other hand, glacier response to a change in climate is rarely instantaneous, mainly because it takes time for an increase or decrease in ice volume in the accumulation zone to be felt at the snout (see [Figure 2.13](#)). Even though changes are transmitted down-valley four times faster than the actual rate of ice movement, there will still be a time lag in glacial response amounting to a few decades for an alpine glacier and a few millennia for a major ice sheet. In any case, portions of the major ice sheets – past and present – have had margins ending in the sea rather than on land, and in these cases, sea level and water temperature have been a more important control over ice-melt than air temperature. Geological evidence of former ice extent comes from moraines and till sequences which can be dated by ^{14}C , dendrochronology and lichenometry. The existence of moraines several kilometres down-valley from modern mountain glacier snouts represents convincing evidence of Holocene fluctuations in climate (Nesje and Dahl, 2003). The most recent moraines found in the European Alps are little more than a century old and have been only partially recolonised by vegetation (see [Plate 2.9](#)). In contrast to those found in mountain valleys, the moraines created by Pleistocene ice sheets now lie far removed from any existing glaciers and are now covered by well-developed soils and vegetation. Some ice sheets, of course, have survived right through the Holocene, notably in Antarctica and Greenland.

Figure 2.13 The glacier–climate linkage.

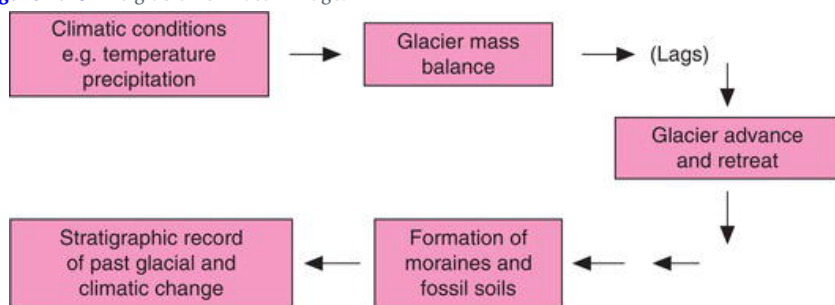
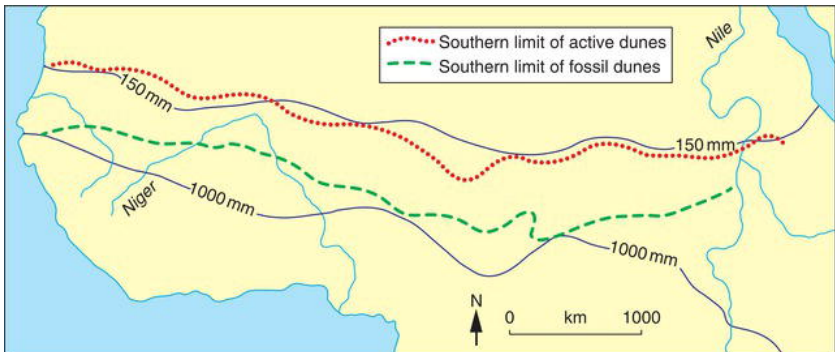


Plate 2.9 Oblique air photograph of the Arolla (left), Pièce (centre) and Tsidjiore Nouve (right) glaciers in Switzerland. The moraines and trimlines visible down-valley of the modern glaciers were created at times during the Holocene when the ice was more extensive and the climate cooler and snowier. The most recent advance occurred during the ‘Little Ice Age’ which ended in the nineteenth century. Photo credit: Swissair.



Figure 2.14 Modern and fossil dune limits on the south side of the Sahara (adapted from Goudie, 1983; adapted from Lancaster, 1979; reproduced with permission of Oxford University Press).



Geomorphology provides important records of Holocene environmental change in the tropics and sub-tropics too, although in this case, the primary climatic variations have been from wet to dry (Douglas and Spencer, 1985). Two of the most climatically sensitive geomorphic regimes are lakes and desert dunes. **Aeolian** processes are most effective where sediment is not bound together by surface soil or vegetation and which consequently has poor internal cohesiveness when dry. Windblown silt, or loess, originating in some cases from deflation of glacial sediments left behind by retreating ice sheets and in others from high-altitude rock weathering in mountains. However, it is in arid lands where ideal conditions for aeolian processes are most often met, with climatically induced

moisture deficiency causing vegetation to be sparse or absent. Because of the close relationship between aeolian processes and climate, sand dune fields help to provide an indication of the changing extent and location of the world's arid zone (Sarnthein, 1978; Lancaster, 1990; Stokes et al., 1997; Hesse et al., 2004; Telfer and Thomas, 2007). On the southern (Sahelian) margin of the Sahara, active sand dunes are today restricted to land with less than 150 mm mean annual rainfall (see [Figure 2.14](#)). Stable, vegetated dune ridges are nonetheless found up to 450 km south of this limit, in areas presently receiving up to 1000 mm rainfall (Grove and Warren, 1968). These mark a former extension of the Sahara southwards, when there was reduction of effective precipitation to only a quarter of its present value. Aeolian landforms not only indicate times of climatic aridity but also reveal former wind strengths and directions (e.g. Wells, 1983; Lancaster et al., 2002; Roskin et al., 2011). Because dunes are aligned relative to the dominant wind vector, they may be used to establish palaeo-wind directions, and differences from the present can then be computed.

If dune activity can indicate periods of climatic aridity, high water levels in **lakes** usually reflect wet phases or **pluvials**. In hydrologically closed or non-outlet lakes, water is lost only through evaporation, assuming there is negligible groundwater outflow (Street, 1980). Lake area and water depth adjust so as to increase or reduce evaporation losses, in the same way as a glacier expands or contracts in response to changes in snowfall and melting. In some cases, closed lakes can behave like giant rain gauges: more rainfall (or less evaporation) and the water level rises, less rainfall (or more evaporation) and the water level falls. Direct geomorphological evidence of former high levels comes from shoreline terraces left 'high and dry' above present-day lakes; indirect evidence comes from proxy records of changing lake salinity (Street-Perrott and Harrison, 1985; Burrough and Thomas, 2009). When a lake moves from being open to closed, it also changes from freshwater to saline (Langbein, 1961). Instead of being washed away down the outflow stream, solutes are retained within the lake water and are progressively concentrated by evaporation. Some closed lakes have become so concentrated that they are now hyper-saline; the Dead Sea, for example, now has a salinity 10 times that of ordinary seawater. Palaeo-salinities are recorded in the chemical composition of lake sediments, because different salts are precipitated as lake salinity increases, in a series from freshwater carbonates through sulphates to chlorides (Teller and Last, 1990). Equally useful in salinity reconstruction are the surviving hard parts of aquatic organisms such as diatoms, ostracods and molluscs (De Deckker, 1981; Gasse et al., 1997). Ostracods can also be analysed chemically for trace elements such as strontium, calcium and magnesium, whose ratios are salinity- and temperature -dependent, and for stable isotopes (Chivas et al., 1986; Von Grafenstein et al., 1999; Holmes and Engstrom, 2003).

Climatic calibration of lake-level data has been attempted using a number of different approaches. One involves the calculation of past precipitation, runoff and evaporation for a lake and its catchment based on a model of the contemporary water balance (Street, 1980). The major assumption that this approach makes is that past temperatures – the primary control over evaporation rates – are known. An alternative approach developed by John Kutzbach (1980) utilises a combined energy and water balance model. Both methods have been used to provide estimates of past precipitation in the tropics (Kutzbach, 1983;

Bergonzini et al., 1997; Rowe and Dunbar, 2004). Other forms of mass balance modelling can involve a lake's salt content or its $\delta^{18}\text{O}$ composition (e.g. Benson and Paillet, 2002; Steinman et al., 2010). Some lakes are mainly fed from below rather than above and act as groundwater 'windows' rather than as rain gauges. In such cases, the whole groundwater system has to be modelled, as was done for the lakes of the Parkers Prairie sandplain in the Midwest United States (Almendinger, 1993).

Geo-archaeology

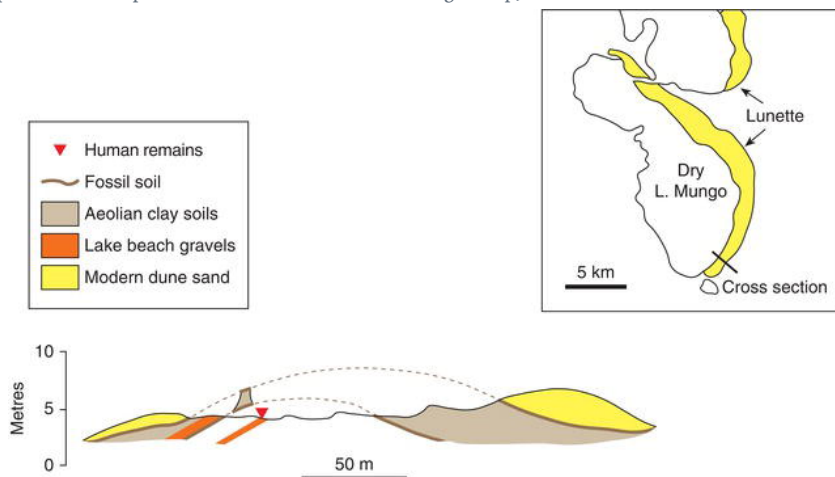
The interface between earth science and past human activity is termed **geo-archaeology** (Davidson and Shackley, 1976; Rapp and Hill, 1998; Goldberg and MacPhail, 2006). It includes the study of sediments from archaeological sites but also extends off-site to investigate, for example, palaeogeographic reconstructions of coastlines of historical importance (e.g. Kraft et al., 1977) and sites buried by river alluvium (Brown, 1997). One of the ways archaeology and geology have helped each other has been in the study of **cave** deposits. Most caves are ancient landforms, and during the Quaternary, they have served as the residence of many creatures, including owls, bears and our own human ancestors. Indeed it is popularly – but wrongly – believed that all early hominins were 'cave men'. As caves have been infilled with sediment, they have also incorporated in their stratigraphy materials brought in by their former occupants: bone remains, stone tools, charcoal from hearths and food refuse. These provide archaeologists with rich pickings for fieldwork, and cave sequences such as at Franchthi in Greece (Farrand, 1999) and Creswell Crags in England (Pike et al., 2005) have been meticulously excavated in collaboration with environmental scientists (e.g. Woodward and Goldberg, 2001), often employing techniques such as micromorphological analysis (Courty et al., 1989).

Interdisciplinary geo-archaeological research has many potential advantages (Vita-Finzi, 1978; Stein and Farrand, 1985). Archaeological and palaeoenvironmental data can be matched and dated from the same stratigraphic context without the need for correlation between sites. In the case of Lake Mungo in Australia, it was survey work on lunette dunes by geomorphologist Jim Bowler that led to discovery of an archaeological site in the first place (Bowler et al., 1972). The human remains found at Mungo pushed back the antiquity of aboriginal peoples in Australia to over 25 ka BP (now more than twice that age; Bowler et al., 2003) and also helped to provide a timescale for the geological and climatic changes recorded in the site's sediment stratigraphy (see [Figure 2.15](#)). In addition, the Mungo skeleton provided an age for the beach gravels with which it was associated and hence for a phase of high lake levels.

Over the course of the Holocene, human impact on geomorphic processes has increased progressively, especially on the rate of **soil loss** from slopes. Soil is eroded and lost as part of the natural geological cycle of denudation, but under dense grassland or woodland that loss is minimised because rainfall is intercepted and runoff reduced (Thornes, 1987). Removal of the vegetation, by human or natural agencies, causes increased potential for erosion by raindrop impact,

surface sheetwash, rilling, gullying, leaching and wind action (Limbrejy, 1975; Boardman et al., 1990). Nor is impact restricted to the site of degradation. Eroded soils are washed downslope and downstream to push up the sediment loads carried by rivers. In fact, soil erosion is an important indicator of the degree of disturbance in an ecosystem.

Figure 2.15 Sediment and archaeological sequence at Lake Mungo, Australia (Bowler et al., 1972; reproduced with permission of the Nature Publishing Group).



Geomorphic environments which record soil erosion histories may be broadly divided into those of net erosion and those of net deposition (Bell, 1983; Bell and Boardman, 1992; Dearing, 1994). The former ought to provide the more accurate soil loss data, but on their own, degraded hillslopes are an unreliable guide to erosion history, their mere existence providing no clue to the age or origin of erosion. However, where archaeological sites lie *in situ*, the former soil surface may be preserved beneath them. Examination of soils beneath prehistoric burial mounds has shown that on the English chalk downlands, later Holocene erosion has almost completely removed the original cover of loess (Catt, 1978).

Most studies of past erosion rates have focused on the point of deposition rather than on the point of erosion. Suitable environments include valley fills, floodplain and estuarine deposits and lake sediments. In all these depositional settings, sediment yield is used as a surrogate for soil loss, an assumption which tends to underestimate true erosion rates. In using sedimentary records, it is best to use relatively small, 'closed' catchments such as drainage basins with a high trap efficiency, often containing a lake or reservoir (e.g. Foster et al., 1985). In order to overcome the problem of differences in the rate of sedimentation across the surface of these 'sinks', it is desirable to examine multiple cores or cut sections. These can be **correlated** with one another by methods such as **magnetic susceptibility**. This makes it possible to establish which places have had fast and which have slow sedimentation rates and to select the most representative sediment profiles for detailed dating. In the case of a lake, accumulation rates can be calculated for different time periods in the past once cores have been dated

and correlated. The mean rate of sediment accumulation over the whole lake bed (e.g. recorded as $\text{gm/cm}^2/\text{year}$) provides a measure of the rate of sediment influx into the lake, which in turn can be converted into mean catchment erosion rate (e.g. t/ha/year) (Dearing, 1986; Foster et al., 2011).

This approach to environmental reconstruction uses sediment yield as a surrogate for soil loss, but it cannot take proper account of material temporarily stored upstream (Trimble, 1999), so it is difficult to calculate absolute differences in erosion rates between catchments. A second problem is that with the exception of non-outlet lakes, even good sediment traps will incur some downstream loss of material, especially of solutes and fine particles such as clay. Thirdly, sedimentary records produce an aggregated record of erosion and give no immediate indication of where erosion took place within the catchment. Fortunately, it may be possible to use magnetic and other properties of the sediment to pinpoint the source areas from which it was derived (Thompson and Oldfield, 1986). Soils, for example, have a different magnetic 'signature' from the parent bedrock because of the formation of secondary ferrimagnetic oxides such as magnetite. In Loe Pool, Cornwall, a layer of red clay recorded in sediment pools was found to be rich in haematite (Fe_2O_3) and tin (Sn). Its metalliferous character could be related to tin mining in the lake catchment during the 1930s (Coard et al., 1983). Although extrapolated catchment erosion rates increased to a mean of c.4 t/ha/year at this time, this would have been almost entirely due to the point source where mine waste was discharged. It is unlikely that agricultural soil losses elsewhere in the catchment were much different from their recent level of 0.2 t/ha/year .

Modelling the past

Models of environmental reconstruction

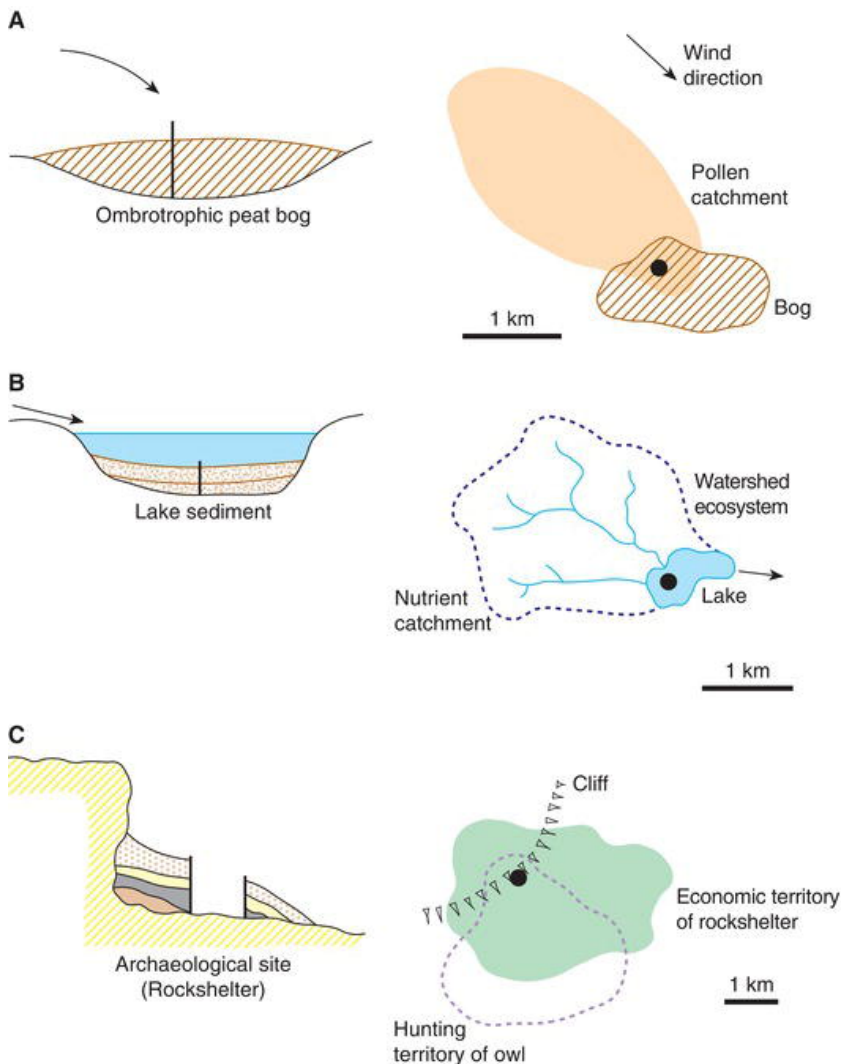
Thus far, palaeoenvironmental techniques have been discussed individually. The question therefore remains of how they and individual site investigations should be fitted together to provide a comprehensive picture of environmental change. One conceptual framework frequently used is a systems model, in which different physical, chemical and biological components of the environment are seen to be interconnected over a definable area of space. A good example is the **lake-catchment ecosystem**, in which a natural and easily defined system boundary is provided by the watershed of the drainage basin (Oldfield, 1977; O'Sullivan, 1979). Nutrients and other materials are washed downslopes into streams and eventually end up in the lake. Not only does the lake therefore serve to integrate changes over the whole catchment – that is, across space – it also records the history of such changes through time. Whereas materials from most drainage basins would be lost from the environmental system along with stream discharge, in a lake catchment, they are trapped and incorporated in bottom sediments, so preserving a record of past environmental processes (Binford et al., 1983; Smol,

2008). Furthermore, a range of dating and analytical techniques can be used in combination on the same lake sediment cores, for example, to reconstruct both catchment erosion rates via sediment influx calculations and a history of land use via the pollen record – the so-called ‘multi-proxy’ approach (Lotter, 2003; Birks and Birks, 2006).

Although palaeoenvironmental data typically derive from specific field localities, the material they contain does not originate solely from just that one place. In other words, each individual site locus actually represents an aggregated record over a wider catchment area. The size and shape of each site catchment will vary according to the type of site and other factors (Vita-Finzi, 1978). Whereas a lake’s catchment is normally the same as its drainage basin, that of a pollen core on a peat bog is defined atmospherically, being the area upwind during the pollen production season (see [Figure 2.16](#)). Birds, animals and human populations may be thought of as having catchments too. The faunal remains found in a cave reflect the area exploited by its occupants, whether a Mesolithic hunting band or an owl. The nature of the territory around an archaeological site helps to explain the on-site record of economic activities preserved in seed and bone remains or in historical archives for individual estates or monasteries (Vita-Finzi, 1978; Butlin and Roberts, 1995).

Of course, catchment areas may not be so easily defined as the examples shown in [Figure 2.16](#). Even so, they serve to emphasise that environmental reconstruction takes place principally at the landscape scale (Birks et al., 1988; Anderson et al., 2006). In order to study environmental processes operating at a smaller scale – say, within a stand of trees – it is necessary to select those sites which only reflect local changes (Bradshaw, 1988). Whereas the pollen entering a large lake normally reflects regional-scale vegetation (c.100 km²), the pollen found in a small woodland hollow will have travelled a short distance and have been determined instead by the nearby flora (c.0.1 ha). At the opposite extreme, large-scale changes involving vegetation formations or climatic zones (c.1 million km²) can only be derived by combining data from many individual sites. For example, pollen records from over 1100 sites have been included in the European Pollen Database (Fyfe et al., 2009) with similar numbers in the equivalent pollen databases for North America, Africa and other continents. These pollen databases can be used to reconstruct continent-wide vegetation changes since the LGM (<http://www.ncdc.noaa.gov/paleo/pollen.html>).

Figure 2.16 Palaeoenvironmental sites and their catchments: A ombrotrophic peat bog; B lake-watershed ecosystem; C rockshelter.



Just as there is a spatial resolution to the reconstruction of past environments, so there is also a temporal resolution. Palaeoecological and geological evidence is resolvable down to one year in the case of varves and tree rings, but more usually the time interval between sample points is 10–100 years. While this inevitably means the 'loss' of information about year-to-year environmental variations, it is compensated by the 'gain' of a smoothed record of underlying trends over centuries and millennia. These are timescales of change which are too long to be observed directly, and for which proxy records are uniquely well suited.

Tying the various complementary sources of palaeoenvironmental data together needs good liaison between different specialists. For this reason, many projects on environmental history involve a team of researchers working over a

number of years at a particular site or in a particular area, whether this be on a long core through an ice cap, or a study of human transformation of a region's landscape. Integrating information from archaeological excavations, historical records, off-site pollen cores, alluvial sequences and dating them all can be a major challenge (not to say headache!), but at its best, the combined results can reveal much more than the sum of their component parts. For example, in the Ystad project, a coordinated team of archaeologists, palaeoecologists, landscape historians and others worked together over a number of years to uncover the history of one area of southern Sweden (Berglund, 1991). There are limitations to any one technique, but together they are able to provide us with an insight into a past that would otherwise remain inaccessible.

Computer model simulations

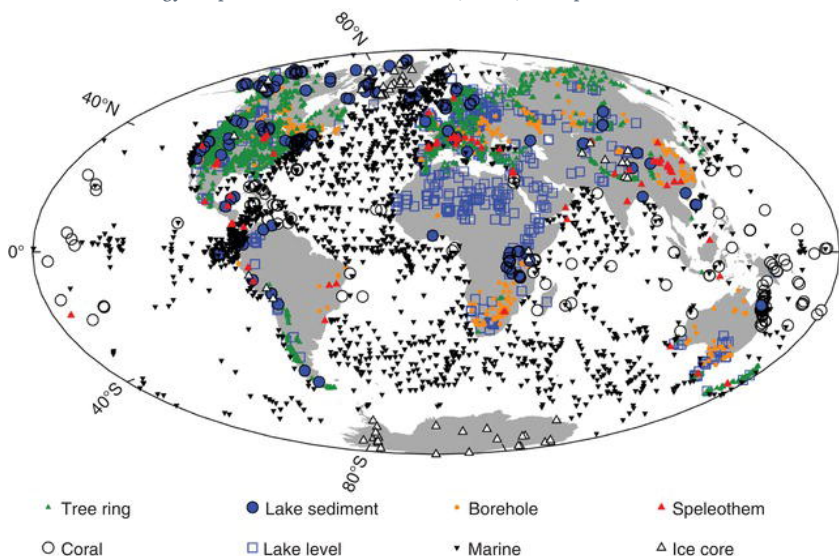
This chapter has described the multitude of different ways in which past climates and environments can be reconstructed. All of these different 'archives' are based on empirical evidence: pollen grains preserved in peat bogs, tree-ring records from ancient timbers, moraines lying down-valley from their modern glaciers, and so forth. However, if we want to scale up these data to look at global as well as local changes, then an alternative, quite different approach becomes increasingly important, namely, the use of computer models to simulate past environments. Rather than starting from the ground and working up, these models represent a 'top-down' approach which try to answer 'what if?' questions. What if a large volume of cold freshwater suddenly emptied into the North Atlantic, for example (cf. Broecker, 1994)? Would it cause the thermohaline oceanic circulation to stop conveying warm water to northwest Europe? The computer models that can be used to investigate these kind of questions start by trying to simulate the relevant natural system today (in this case the ocean circulation), albeit in a much simplified form. Once the computer simulation matches the pattern seen in the real world, then one or more of the controlling parameters can be changed in the computer model (in this case, freshwater influx), and the resulting changes to the simulation recorded. Models have the great advantage of permitting predictions as to how an environmental system might change through time. For this reason, they have played a central role in the scientific work predicting future climate change related to the increasing human-induced emissions of greenhouse gases (Hegerl et al., 2007).

Of course, all models are wrong by definition, because even the most complex and sophisticated of them represent simplifications of reality. What matters is whether a model is 'wrong' by 5% or by 50%. While the starting point is to ensure a good fit between the results from a model under present-day conditions, this in itself is not sufficient. Equally important is how successfully a model captures changes to a system. For this reason, one of the main ways to test the accuracy of computer models is against conditions different from those prevailing today. Because – by definition – the future is still unknown, these different scenarios can only come from times in the past, and the most important of them come from the period since the **LGM**. A variety of models have been used to simulate the climates of the past 20 000 years, ranging from simple energy balance models to

the computer-intensive models of the atmosphere–ocean–sea ice–land surface that are also used to project future climate (Trenberth and Otto-Bliesner, 2003; Valdes, 2003). Comparing model results against reconstructions based on proxy data requires that pollen, foraminifera, lake-level and other data are compiled and organized in a standardized form with good geographical coverage (Figure 2.17). The first systematic, global-scale attempt at model: data comparison was by the CLIMAP Project Members (1976) and used only marine data. This was followed by the COHMAP Members (1988; Wright et al., 1993), which used proxy climate data from both land and oceans. This in turn was succeeded by PMIP, which has compared the experimental results between different computer models, as well as against proxy data (e.g. Pinot et al., 1999; Braconnet et al., 2004). The outcomes from these and other model : data intercomparison projects are discussed later in this book, but suffice it to say that they represent a powerful way of advancing our understanding of past – and future – global change. When the data and models agree, it gives confidence in our reconstructions. When they do not, it indicates that something is still missing in the simulations, or that our proxy data are incomplete.

While most attention has focused on the global climate system, computer model simulations have also been made of the oceans, ice sheets, the Earth's crustal dynamics, the carbon cycle and global land vegetation (e.g. Prentice et al., 1992; Mitrovica and Vermeersen, 2002), which have the potential to be coupled together to create an Earth system model. Similarly, proxy data sets now exist not only for past climate and vegetation (Figure 2.17) but also for other components of the global environment such as atmospheric dust (DIRTMAP; Kohfeld and Harrison, 2001) and wildfire (via microcharcoals; Power et al., 2008).

Figure 2.17 Global distribution map for different records in the NOAA/NCDC World Data Center-A for Paleoclimatology. Reproduced from Evans et al., 2013, with permission of Elsevier.



Conclusion

Only selected techniques of dating, environmental reconstruction and modelling have been reviewed in this chapter. Fortunately, several texts provide further detail on these and other palaeoenvironmental methods, for example, by Berglund (1986), Bradley (1999), Lowe and Walker (1997) and Mackay et al. (2003).

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The Pleistocene prelude ($> 11\,700$ Cal. yr BP)

Ice Age environments

If we draw back in time from the start of the Holocene, some 11 700 years ago (11.7 ka BP), it soon becomes clear that our present epoch is but the latest of a series of warmer intervals that have punctuated the otherwise chilly climate of the Quaternary. The last few million years have been marked by frequent climatic changes, with repeated oscillations between colder and warmer conditions. Warmer intervals lasting more than about 10 000 years are known as interglacials, while the colder periods are termed glacials. During the latter, massive ice sheets developed over Scandinavia and Canada, only to melt away entirely during the interglacials. Interspersed within glacials and interglacials were cold and warm intervals of shorter duration (e.g. 1000 years), known as *stadials* and *interstadials*, respectively.

The glacial–interglacial cycle

For many years, a scheme involving four *Pleistocene* glacial-interglacial cycles was recognized in many land areas. The evidence for this view of the Quaternary Ice Age originally came from moraines, erratic boulders, striated and polished rocks in areas such as the Alpine Foreland of Switzerland and southern Germany. Since 1950, more complete records of Pleistocene climatic change have been available from the study of deep-sea sediment cores, in particular using oxygen isotope analysis (see *Technical Box III*). Deep-sea cores provide unbroken stratigraphic records and show that schemes like the four-glacial model established by Penck and Bruckner for the Alps underestimated the true number. About eight major glacial–interglacial cycles have in fact occurred during the past 0.8 million years, each cycle lasting c.100 000 years, and with other cycles occurring before this time (Williams et al., 1998). The deep-sea record of past climate change now has its counterpart on land from cores drilled through the East Antarctic ice sheet (EPICA Community Members, 2004), thick loess deposits

in central Europe and Asia (Porter and Zhisheng, 1995) and long pollen records from sites in Greece, the Colombian Andes and elsewhere (Hooghiemstra et al., 1993; Tzedakis, 1993).

TECHNICAL BOX

Technical Box III: Dividing up late Quaternary time

The Quaternary Ice Age is conventionally divided up into periods of colder (glacial) or warmer (interglacial) climate. Initially these glacial and interglacial stages were defined using land-based evidence from Europe and North America, parts of which were formerly covered by ice sheets but are now ice-free. Different stage names were used for each region. Thus, the last glaciation in North America is called the Wisconsinan, it is the Würm in the Alps, the Weichselian in northern Europe and the Devensian in Britain. Further back in time, uncertainties increase about the completeness of the stratigraphic record in each of these regions, and about correlations between them. Consequently, since the 1970s, these schemes have been replaced as a global template by the oxygen-isotope ($\delta^{18}\text{O}$) record from deep-sea cores (Imbrie and Imbrie, 1979; Lowe and Walker, 1997; Williams et al., 1998). This has the great advantage of being complete and unbroken through time. And, because the deep-sea benthic $\delta^{18}\text{O}$ signal reflects globally averaged ice volume, it should also – in principle – be the same anywhere in the world.

A scheme for labelling Marine Isotopic Stages (MIS) was initially developed by Cesare Emiliani. In this, he gave warm stages odd numbers (1, 3, 5, etc.) and cold stages even ones (2, 4, 6, etc.), counting back from the present, with the current Holocene stage as MIS 1 (see Figure 3.2). On that basis, the last

interglacial stage ought to be MIS 3, except that it is not! Emiliani instead erroneously decided to label a warmer interval within the last cold glacial stage as MIS 3, with the **LGM** as MIS 2 and the early part of the last (Wisconsinan, Weichselian, etc.) glaciation as MIS 4. The last interglacial is consequently labelled MIS 5. Other, earlier climate stages do fit the expected pattern, however (i.e. MIS 7, 9, 11, etc.). Some of these major isotope stages have themselves been sub-divided. The last interglacial complex, for example, had three warm peaks labelled respectively MIS 5a, 5c and 5e (MIS5e being the last interglacial proper), separated by two cold troughs labelled MIS 5b and 5d. In addition, shorter-lived events have been given names such as the Younger Dryas stadial or the Heinrich 1 event, although these were not global in extent.

Evidence for interglacial conditions comes from fossil soils, buried peat and lake sediments and other deposits containing warm-climate faunas. The last true interglacial, known in continental Europe as the Eemian, in Britain as the Ipswichian and by other names elsewhere, began around 130 000 years ago and lasted for around 14 000 years. During at least part of that time, it appears that climatic conditions were slightly warmer than those which have been experienced during the Holocene. In consequence, the volume of the Greenland ice sheet was reduced to only about half that of today (Overpeck et al., 2006), and sea levels were up to about 5 m higher than during recent millennia (Cuffey and Marshall, 2000; Siddall et al., 2008). Earlier, interglacials also seem to have been broadly similar in character to the Holocene, although some, such as MIS 11 around 0.4 million years ago, lasted longer, up to 26 000 years (EPICA Community Members, 2004). In one respect, however, the Holocene stands in sharp contrast with the Pleistocene interglacials, namely the mark of ‘man’ imposes itself far more dramatically on the present warm epoch than on any previous one. The emergence and evolution of the human race is one of the most important distinguishing characteristics of the Quaternary, but anatomically modern humans (*Homo sapiens* var. *sapiens*) only emerged 100–200 000 years ago. The main radiation took place during the last glacial period, with Neanderthal populations being replaced in western Europe between about 45 000 and 25 000 years ago (Finlayson, 2004; Finlayson et al., 2006; Mellars, 2006). During most Pleistocene interglacials, only ancestral humans were present, and in many parts of the globe, such as the Americas, even they were absent.

Given that the impact of early hominids on the natural world was rather

slight, it is possible to use Pleistocene interglacials as analogues for what the Holocene world would have been like had it not been disturbed by human agency (Ruddiman, 2005). Pollen analysis of peat and lake deposits has revealed the vegetation changes that took place during Pleistocene interglacial periods (Davis, 1976). Although there were some floristic differences between interglacials, they were minor in comparison with their overall similarity. This is all the more remarkable considering that many plant formations had to be recreated from scratch at the beginning of each warm interval. In northern Europe, Johannes Iversen (1958) proposed a simple phase model for interglacial vegetation development. The pre-temperate **protocratic** phase was one of immigration of tree species from southern refuges, and the first arrivals were normally boreal taxa such as birch (*Betula*) and pine (*Pinus*). During the subsequent **mesocratic** phase, mixed deciduous forest became established (see Table 3.1). Shade-giving trees such as oak (*Quercus*) and elm (*Ulmus*) replaced the pioneering light-demanding genera. Although the mesocratic phase was usually of longer duration than the protocratic one, it was far from static floristically. Pollen diagrams show that new trees continued to arrive throughout the duration of interglacials, with floristic diversity progressively increasing as a result (Davis, 1976).

Table 3.1 Ecological characteristics of trees associated with different phases of the northern European interglacial cycle
Source: Birks, 1986. Reproduced with permission of John Wiley & Sons, Ltd.

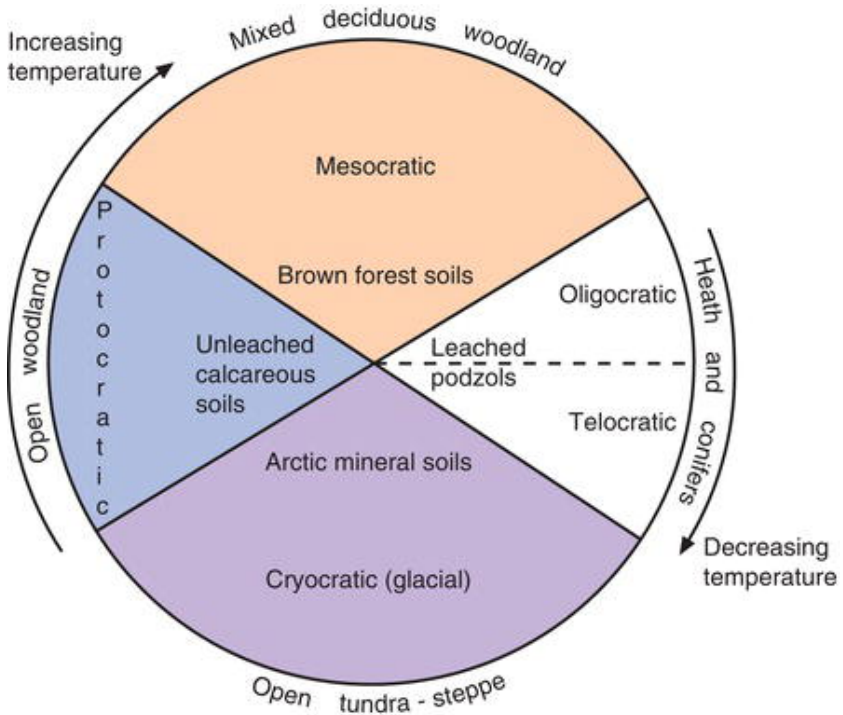
	Protocratic	Mesocratic	Oligocratic
Typical tree taxa	Birch	Oak	Beech
	Aspen	Elm	Spruce
Age of first seed setting	Young	Mature	Mature
Seedling tolerance to shade	Intolerant	Tolerant	Tolerant
Dispersal efficiency	Good	Poor	Poor
Migration rate (m/year)	>1000	500–1000	<500
Growth rate	Fast	Slow	Slow
Longevity	Short	Long	Long
Soil preference	Fertile	Brown earth	Podzol
	Unleached	Mull humus	Mor humus
Ecological traits	Ruderal	Competitive	Stress tolerant

Soil changes also influenced the composition of the temperate forests, especially in areas that had been subject to glaciation during cold periods of the Pleistocene (see Figure 3.1). Here, leaching of weathered bases from glacial tills led to a shift from neutral to acid soils, favouring trees like spruce (*Picea*) which mark the **oligocratic** phase of an interglacial. Once established, conifers produced further soil acidification as their fallen needles created an acid mor humus. The final retrogressive **telocratic** phase saw the replacement of mixed deciduous forest by heathland and open coniferous woods, particularly where soils were locally prone to leaching out of their nutrients. Thermophilous trees disappeared as the climate changed from interglacial to glacial, but the order of departure was

not simply that of arrival in reverse. It seems as if many deciduous trees did not have time even to ‘pack their trunks’ and head off towards the Mediterranean sun at the end of the last interglacial (Woillard, 1979; Bennett et al., 1991).

Most high-latitude pollen records stop with the switch from interglacial to glacial conditions. On the other hand, this is not so with sites from the tropics and subtropics (Metcalfe and Nash, 2012). They show that vegetation formations like the rainforest of northeastern Australia and the xerophytic woodlands of the Mediterranean Basin and were virtually eliminated during glacial phases of the Pleistocene (Kershaw, 1986; Tzedakis, 1993). The tropical rainforests of Africa and Amazonia were also reduced to isolated pockets at these times (Mayle et al., 2004). Climatically induced vegetation change was consequently not restricted to the northern forests but affected many of the world’s biomes. What is more, this great natural deforestation of the Earth was repeated many times during the course of the Pleistocene.

Figure 3.1 The interglacial cycle in northern Europe, for example, Denmark (adapted from Iversen, 1958).



In short, the million years or so prior to the Holocene was a period of extraordinary environmental variability and stress upon ecosystems and organisms. Those species capable of adapting to a fluctuating environment were able to benefit. Our own species is a prime example, for Pleistocene environmental change undoubtedly helped speed the pace of human evolution (Roberts, 1984). There were unusually high rates of speciation and extinction in

many areas of the world during the Pleistocene. In South America, for instance, repeated fragmentation and coalescence of the rainforest encouraged the creation of new species and may be linked to the high species diversity of that biome (Prance, 1982; Colinvaux, 1987; Colinvaux et al., 2000). In most cases, however, evolutionary changes brought about by Pleistocene fluctuations in climate were secondary to those which occurred in species distribution and population numbers. During the Pleistocene, species shifted their geographic distributions, with many experiencing alternate fragmentation and reconstitution of their ranges, while others simply tracked their habitats by moving towards or away from the poles, or up and down mountains. The glacial–interglacial range shifts made by creatures ranging from plants and insects to mammals frequently involved distances of several thousands of kilometres (Bennett, 1997). Soil- and land-forming processes were subject to a similar intensity and rapidity of change during glacial–interglacial cycles. This was most acute in areas which were glaciated. The soils of the Canadian Shield were literally scraped up and dumped in the American Midwest every time the Laurentide ice sheet advanced southwards – a later gift from Canada to US farmers.

Understanding the causes of long-term climatic change

It is now widely accepted that the main trigger for Ice Age climatic fluctuations was small variations in the Earth's orbit around the sun. This idea goes back to nineteenth-century scientists such as Adhémar and Croll, although a coherent astronomical theory of climate change was not put together until the 1920s. This was the work of Serbian mathematician Milutin Milankovitch who integrated three astronomical cycles of different length (Table 3.2), which were superimposed on each other to produce a more complex pattern of change through time (Berger, 1992). Only one of these, the eccentricity cycle, caused the Earth as a whole to receive different amounts of solar radiation. The other two factors (tilt and precession) involved its redistribution between seasons in the different hemispheres. On that basis, we might have expected glacial phases to have taken place alternately in the northern and the southern hemispheres. In reality, the overall pattern of glacial and interglacial stages was the same across the globe. It seems, therefore, that the effects of orbital variations over the northern hemisphere land masses triggered mechanisms which acted to cool or warm the planet as a whole. Comparison of Milankovitch's predictions with the deep-sea record of past ice sheet expansion and contraction has shown a remarkably close fit – prompting claims that the mystery of Ice Age changes in climate had been solved (Imbrie and Imbrie, 1979). Certainly, the astronomical theory has proved invaluable, not least because it allows us to make predictions about the future long-term course of climate. It is clear, for example, that we are in the latter part of an interglacial period and that if allowed to run its course naturally, the Earth's climate would start back towards a glacial phase, perhaps some 1500 years from now (Tzedakis et al., 2012).

On the other hand, changes in the Earth's receipt of solar radiation due to

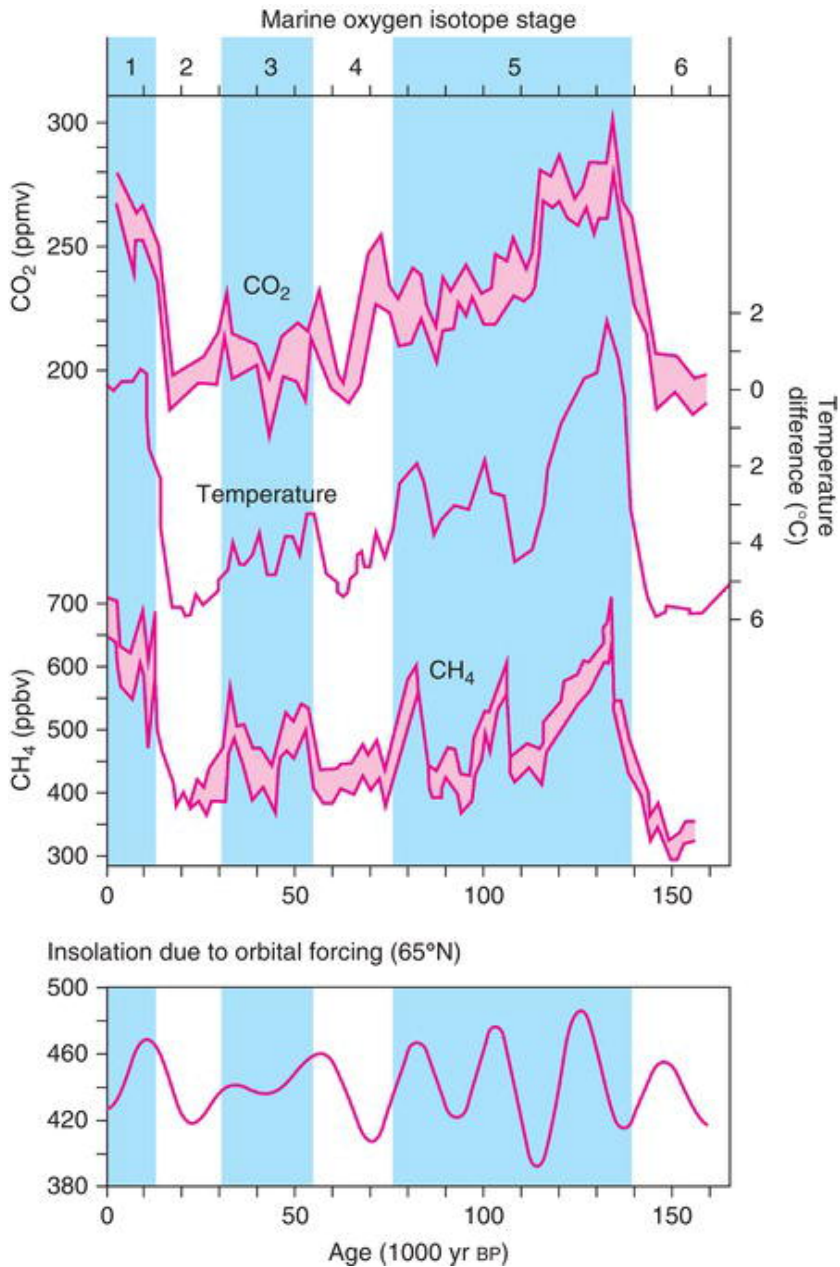
orbital variations were relatively modest – sufficient to cause direct warming or cooling of no more than 2°C. In reality, temperatures fluctuated by much more than this amount in the past, especially on land. For this reason, Jim Hays and colleagues (1976) described variations in the Earth’s orbit as the ‘pacemaker’ of the Ice Age climate. Other forces must have been at work to amplify its signal, most of which would have been internal to the Earth’s own system. One important set of feedbacks involves the linkages between ice, ocean and climate. Many past or present ice sheets, for example, terminate in the oceans, and seawater temperature and sea level, rather than local climate, control how far out the ice is able to grow. A minor increase in melting of the ice triggered by orbital changes would raise sea levels. This in turn would undercut more ice (often by destabilising the floating ice shelves which protect the ice sheet margins from wave attack), and the ice calved off into the sea would raise sea levels still further. This self-reinforcing process would continue until the ice retreated to the point where it was grounded on land, not in the ocean, or until some other factors came into operation to stop the positive feedback loop. Other feedback links between ice, ocean and climate involve surface albedo effects and the impact of cold, fresh meltwater on ocean circulation and on evaporation rates from the ocean surface. In these ways, a small externally driven initial warming could have been amplified to push the climate from a glacial to an interglacial state, or vice versa. It seems that the Earth’s Ice Age climate alternated between two more or less stable states (Broecker et al., 1985).

Table 3.2 The main astronomical cycles of climatic change

	Cycle length (Years)
Orbital eccentricity	100 000
Axial tilt (or obliquity)	41 000
Precession of the equinoxes	> 21 000

Another very important process which amplified the orbital climatic signal was changes in greenhouse-gas concentrations. Gases such as carbon dioxide and methane help ‘trap’ heat within the atmosphere by intercepting outgoing long-wave radiation, and higher atmospheric concentrations would have been associated with enhanced warming of the climate. Estimates of past fluctuations in greenhouse-gas concentrations come from sources such the stable carbon isotope ($\delta^{13}\text{C}$) ratios in peats, lake sediments and soils (Figge and White, 1995; Street-Perrott et al., 1997; Mayle et al., 2004) and the density of stomata on fossil leaves (Beerling et al., 1995). However, our most informative source of information comes from ancient air bubbles trapped in ice sheets and recovered in ice cores. The long Vostok and Dome C (EPICA) cores from Antarctica provide our most complete and direct record of changes in greenhouse gases and temperatures, spanning several glacial–interglacial cycles. As [Figure 3.2](#) shows, during interglacials, atmospheric CO₂ levels lay around 270–280 ppmv, while under full glacial conditions, they fell to 190–200 ppmv. The close match between the curves for CO₂, CH₄ and temperature strongly supports a link between greenhouse gases and past climate change, the former appearing to act as an amplifier of orbital insolation changes.

Figure 3.2 Variations in temperature and greenhouse gas concentrations through the last full glacial-interglacial cycle in the ice core from Vostok, Antarctica (Lorius et al., 1990; reproduced with permission of the Nature Publishing Group), along with variations in solar insolation due to orbital forcing ($\text{Cal/cm}^2/\text{day}$) (data from Berger, 1992).

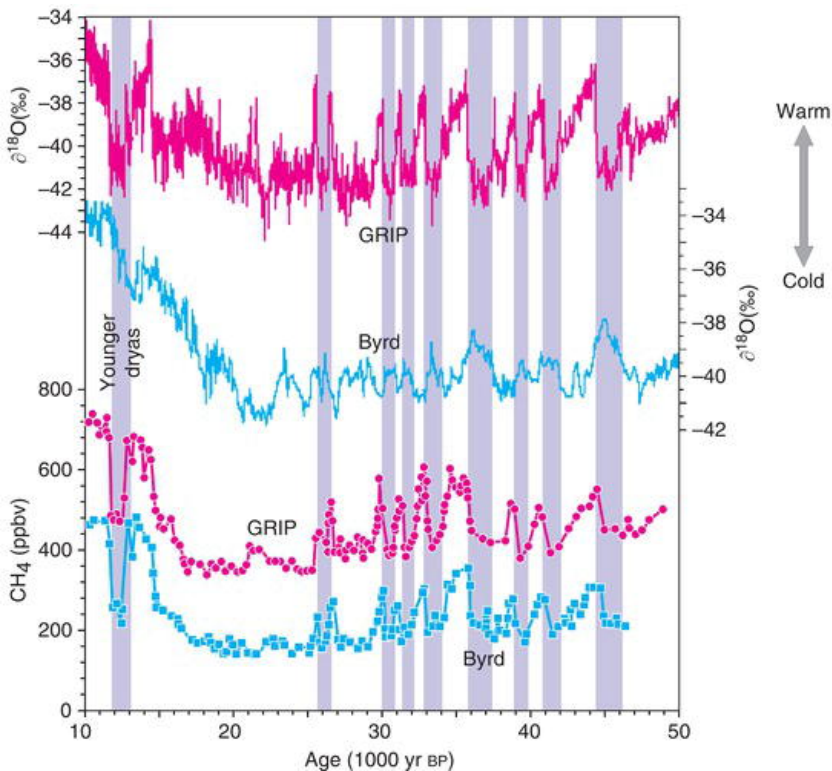


Antarctic ice cores and most deep-sea sediment cores show only modest

climate variability on timescales shorter than Milankovitch cycles. On the other hand, the Greenland ice cores tell a different story. This is partly due to the much faster rate of snow and ice accumulation here, which preserves a signal that has not been ‘smoothed’, but the difference also reflects the rather unique palaeogeography of the North Atlantic, previously surrounded by three separate ice sheets. The first Greenland ice cores, taken in the 1960s at Camp Century, showed very high $\delta^{18}\text{O}$ variability in the section dating to the last glacial stage, but because this core was taken near ice margin, Willi Dansgaard and his co-workers assumed that these short-lived fluctuations were simply ‘artefacts’ created by distortions in the ice as it flowed outwards. However, when cores were taken at the Greenland ice summit 25 years later, both GRIP and GISP2 cores showed the same rapid oscillations, flickering on and off like a faulty switch (Alley, 2000). These Dansgaard–Oeschger (or D-O) oscillations, named after Dansgaard and isotope scientist Hans Oeschger, took place principally during MIS 3 at roughly 1500-year intervals (Figure 3.3). Each one started with an extremely rapid warming, often of an amplitude ($\sim 10^\circ\text{C}$) not much less than the temperature difference between glacial and interglacial climates, followed by a more gradually cooling.

This instability in the glacial climate was not restricted to the Greenland ice cap and has its counterparts in many other high-resolution palaeoclimate records on land and in the oceans spanning the same time period. They include $\delta^{18}\text{O}$ and palaeotemperature changes in cores from the Santa Barbara Basin off the coast of southern California, whose waters became anoxic and led to the preservation of varved sediments (Hendy et al., 2002), and pollen records from both marine and lake cores around the Mediterranean (Allen et al., 1999; Sánchez Goñi et al., 2002). In addition, North Atlantic deep-sea cores show separate evidence of periods of major, concentrated iceberg discharges from the northern and eastern edge of Laurentide ice sheet into the North Atlantic on half a dozen occasions during the last glacial stage – the so-called Heinrich events (Heinrich, 1988; Broecker, 1994). These peaks in ice-rafted sediment debris (IRD) seem to have occurred when the ice sheet went from a state of being frozen to the bedrock to one in which the bed was above pressure-melting point (Bond and Lotti, 1995). On a lubricated bed, the ice thinned and surged outwards, much of into the sea to create a vast armada of icebergs. In some cases, these may have been linked to catastrophic drainage of pro-glacial lakes abutting the Laurentide ice sheet, such as glacial Lake Agassiz (Teller et al., 2002). The Heinrich events appear to have coincided in time with some of the deepest cold D-O troughs, with the last one (the Heinrich 1 (H1) event) occurring immediately before the Late-Glacial climatic warming.

Figure 3.3 Asynchronous climatic variations in Greenland (GRIP) and Antarctica (Byrd) during the last glacial period, as shown by different phasing of the temperature proxy $\delta^{18}\text{O}$ (Oxygen isotopes) compared to methane (modified from Blunier and Brook, 2001). Shading indicates northern hemisphere cold (stadial) phases. Reproduced with permission of AAAS.



The Last Glacial Maximum and after

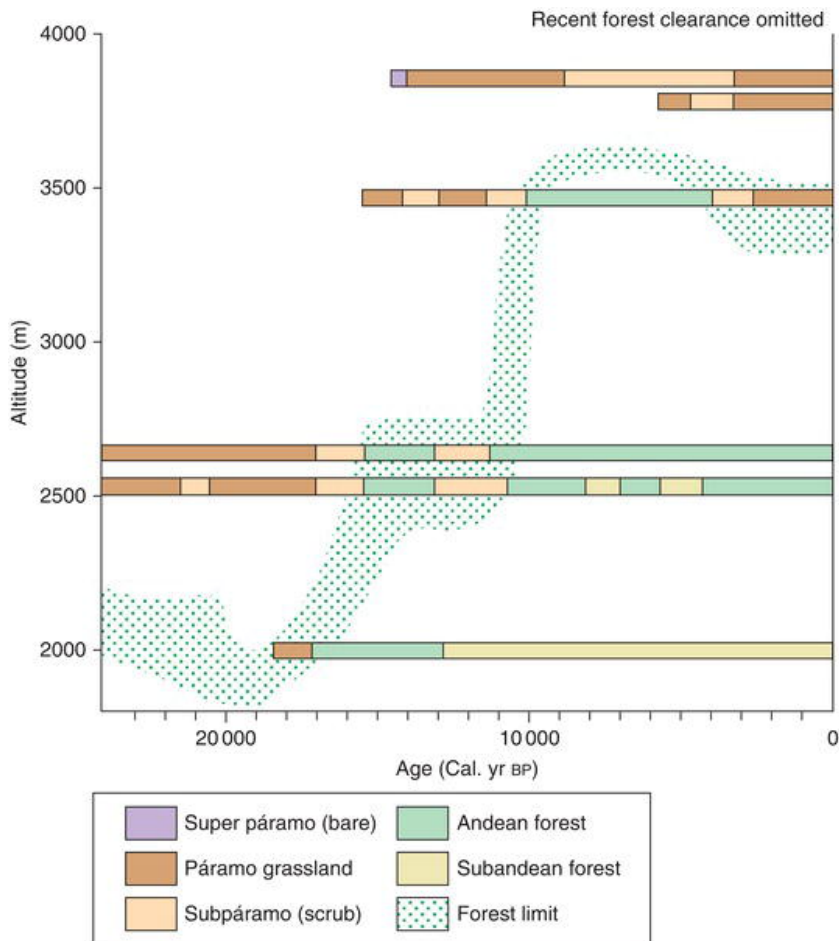
The last glacial period reached a peak between 25 000 and 18 000 years ago. Evidence for what the world was like at this time is relatively abundant and well preserved in comparison with earlier glacial periods whose record has been greatly disturbed by later ice action and other geomorphic processes. Around 21 000 Cal. yr. BP, temperatures on land had fallen by as much as 20°C (Wright et al., 1993), although ocean temperatures changed much less. The reconstruction by the CLIMAP Project Members (1976) indicated that for tropical parts of the ocean, sea-surface temperatures for the Last Glacial Maximum (LGM) were within 2°C of the present, although in some areas, the true temperature fall may have been nearer to 5°C (Guilderson et al., 1994; MARGO Project Members, 2009). Under the full glacial climate, ice sheets over 4 km thick lay over northern Europe and North America, and smaller ice caps existed in the Alps, the southern Andes and parts of East Asia (Denton and Hughes, 1981). Most of northern Asia was too dry for ice sheets to form, and Siberia consequently experienced intense periglacial activity. Much of today's deep permafrost layer is, in fact, a relict from the last glacial period.

The build-up and decay of ice sheets locked up and then released water from

the hydrological cycle, causing ocean volume to fluctuate and sea levels to rise and fall (Peltier, 1994; Siddall et al., 2003). Global sea levels were lowered on average by more than 100 m at glacial maxima, creating new continental land masses in areas such as southeast Asia. Land bridges were exposed, so that it became possible to walk from northeast Asia to Alaska or from Britain to the European mainland. This undoubtedly aided the spread of fauna and of early human populations into new or marginal areas. It was the era of the woolly mammoth and of the Ice Age hunters who portrayed their prey so vividly on the cave walls at Lascaux in southwest France. Most of northern Europe and Asia that was not glaciated formed a tundra-like landscape at this time (Tarasov et al., 2000). Some shrubs such as dwarf birch (*Betula nana*) may have survived the harsh climate but vegetation was sparse and was often dominated by steppic plants. Open-habitat species predominated in most of southern Europe as well, and trees were restricted to isolated refuges in Iberia and southeast Europe (Bennett et al., 1991). On the other hand, this was not true of all mid-latitude regions. Japan almost certainly remained forested through the last glaciation (Tsukada, 1983; Nakagawa et al., 2012), while much of the eastern United States was covered by pine and spruce woodland (Webb et al., 1993). A tundra zone did exist close to the Laurentide ice sheet, but in contrast to northern Europe, it was only 100–200 km broad (Watts, 1983).

In the tropics, temperature depression caused vegetation zones on the equatorial mountains to lower by 1000–2000 m (see [Figure 3.4](#)), but a bigger change was brought about by glacial aridity (Goudie, 1983; Bonnefille et al., 1990; Gasse, 2000; Metcalfe and Nash, 2012). During and after the LGM, most low-latitude areas were significantly drier than at present. This is indicated by landforms and sedimentary records, notably as low lake levels (Street-Perrott et al., 1985; Barker and Gasse, 2003; Hodell et al., 2008) and fossil sand dunes (Sarnthein, 1978; Stokes et al., 1997), the latter extending way beyond their present isohyet limits (see [Figure 2.14](#)). The Indian monsoon was suppressed, reducing the transfer of water vapour from the oceans to the continents. However, there were some exceptions to this widespread aridity, including the rainforest of southeast Asia, parts of which seem to have survived intact through glacial times (Wurster et al., 2010). Another relatively well-watered area was the American Southwest, where a complex system of cascading lakes occupied the now parched deserts of the Owens River and Death Valley prior to 11 000 years ago (Smith and Street-Perrott, 1983). Packrat midden studies have revealed that conifer woodland then occupied what is today desert scrub ([Figure 2.11](#)).

Figure 3.4 Altitudinal shifts of vegetation belts in the Colombian Andes during the late Quaternary, inferred from six pollen sites (after Flenley, 1979). Reproduced with permission of SAGE.



The cold and general aridity of glacial times was not conducive to plant growth, particularly for trees, and led to expansion of savannah grassland, tundra, steppe and other open types of vegetation at the expense of forest biomes (Table 3.3). This in turn led to significantly less carbon being stored in land vegetation and soils. One study (Adams et al., 1990) estimated that this organic carbon store was less than half that in the modern 'natural' biosphere, but other calculations indicate that the true reduction at the LGM was more likely to have been around 500×10^{15} g, a decrease of about one-fifth (Prentice et al., 1993; Van Campo et al., 1993). This presents us with something of a conundrum. Ice cores show that there was about 30% less carbon in the glacial atmosphere, but far from being taken up by land plants and soils, this carbon store was also depleted. So where did the 'missing' carbon go in the Ice Age Earth? The answer would appear to lie in the oceans (Pederson et al., 2003). Some of the removal of carbon can be accounted for by more efficient 'export' of organic matter from the upper photic layer of the oceans to the seabed where it is buried in sediments. But most of it

seems likely to have been caused by a real increase in the productivity of marine plankton.

Table 3.3 The extent of major terrestrial biomes at the Last Glacial Maximum and the late Holocene (excluding agricultural clearance)

Source: Adams et al. (1990). Reproduced with permission of the Nature Publishing Group.

	LGM	Holocene
Tropical forest	11.9	26.1
Temperate forest	7.1	33.7
(Total forest)	(19.0)	(59.8)
Savanna and scrub	44.0	31.1
Temperate grassland/steppe	11.3	15.5
Desert	29.5	17.2
Tundra and polar desert	27.1	9.0
(Total non-forest)	(111.9)	(72.8)
Other (mainly ice)	33.4	17.4
Total	165.3	150.1

Areas in $\times 10^6$ km²

Most open oceans are biological deserts, their low productivity being caused by a shortage of key nutrients such as phosphorus and nitrogen. In many areas of the ocean, however, the nutrient which limits plankton growth is iron (Martin and Fitzwater, 1988), notwithstanding the fact that this is one of the commonest elements of the Earth’s lithosphere. Iron both controls biological productivity directly and also helps to regulate the amount of nitrogen available to plankton (Falkowski, 1997). It reaches the open ocean mainly in the form of wind-blown dust, carried from the continents. The flux of atmospheric dust depends on conditions in dryland source areas, and during glacial times, it was a good deal drier and windier too. This is evident in the record of lake levels, loess and sand dunes in key dust-source areas such as Patagonia, which lies upwind of the vast Southern Ocean (Haberzettl et al., 2009); it is also recorded in greatly elevated dust concentrations in the Greenland and Antarctic ice cores (Lambert et al., 2008) (see [Figure 3.6C](#)). Iron limitation on productivity in many open ocean areas therefore appears to have been eased during glacial periods by a more than fivefold increase in the dust supply carried from the continents. Ironically, the same dry climatic conditions which *decreased* biomass on land appears to have *increased* it in the oceans, and in so doing contributed to lowering the concentration of carbon dioxide in the glacial atmosphere (Kumar et al., 1995; Bopp et al., 2003; Ridgeway, 2003).

After about 18 ka BP, the world’s major ice sheets retreated and thinned. Deglaciation, which took about 9000 years to complete in Europe and North America, may have had as much to do with climatic aridity as with warming. The

very bulk of the ice sheets created anticyclonic conditions and deflected moisture-bearing winds away from the ice sheets' accumulation zone (Wells, 1983). Evaporation from the ocean surface was also discouraged by a cool, low-salinity meltwater layer notably in the North Atlantic (Ruddiman and McIntyre, 1981), especially during periods of major iceberg discharge, for example, during the H1 event around 16–17 000 years ago. The ice sheets therefore appear to have been starved to death by inadequate nourishment of snowfall. As the ice melted and sea levels started to rise, so some ice sheet margins became vulnerable to attack by marine action, especially in areas such as Baffin Sound (Carlson and Winsor, 2012). Great blocks were undercut by the waves at the ice sheet margins to float away as icebergs and eventually raise sea levels still further. The southern, land-based margin of the Laurentide ice sheet retreated more gradually during the early stages of deglaciation, and in consequence, the ice became thinner rather than smaller in extent.

Deglaciation was an unsteady process, and at various times towards the tail end of the Pleistocene, the ice sheets temporarily re-advanced. These glacial re-advances do not appear to have occurred synchronously everywhere. The Laurentide ice sheet even expanded on some sectors of its southern front, while other parts were stationary or retreating. The Green Bay, Lake Michigan and other ice lobes of the Great Lakes region may owe their origin to glacial surging rather than to climatic forcing. However, in many parts of the world, environmental changes occurred during the terminal Pleistocene that were undoubtedly caused by a sharp oscillation in climate, which represented a false start to the present (Holocene) interglacial.

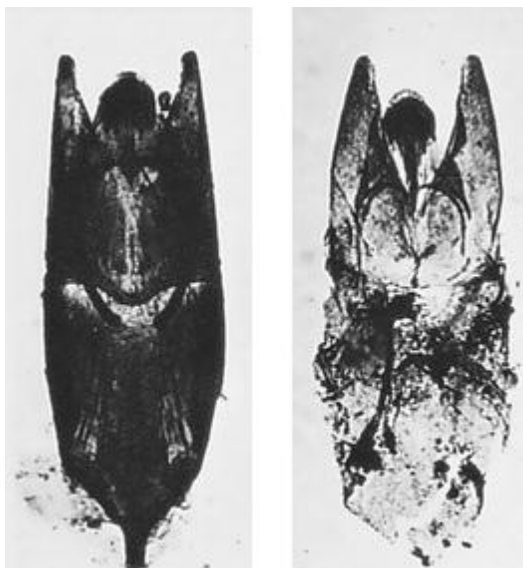
The terminal Pleistocene (15 000–11 700 Cal. yr BP)

The Late Glacial in the North Atlantic region

Around the North Atlantic Ocean, the late Pleistocene thermal oscillation comprised an interstadial (c.15–13 ka BP) followed by a stadial (13–11.7 ka BP). Some of the best evidence for this oscillation comes from the remains of invertebrates, especially beetles and chironomids, which provide the next best thing to a Late-Glacial thermometer ([Plate 3.1](#)). Studies of Coleoptera at a variety of European sites allowed Russell Coope (Coope, 1975; Coope and Lemdahl, 1995) to estimate average July temperature changes from before 15 ka to about 10.5 ka BP. At the beginning of this period, temperate assemblages replaced arctic ones, and at Glanllynau in north Wales, the rate of climatic amelioration is calculated to have been an amazingly rapid 1°C or more per decade. After rising

abruptly soon after 15 ka BP, summer temperatures lay at – or even above – those of today, but Coleoptera suggest the climate then cooled progressively from the optimum of this ‘Windermere’ interstadial (Lowe et al., 1995), equivalent to the Bølling–Allerød events of Scandinavia (Table 3.4).

Plate 3.1 Modern and fossil specimens of beetle genitalia, c.0.8 mm in length, genus *Helophorus*. In some cases, the size of the male sexual organ is the only way to distinguish cold-loving Siberian species from temperate European ones; remarkably these genitalia of Late-Glacial age have been found preserved. Photo credit: *Philosophical Transactions of the Royal Society of London* (B.265, pp. 298–326), R.B. Angus, 1973. Reproduced with permission of the Royal Society of London and Robert Angus.



Trees were slower than invertebrates in responding to the climatic improvement at the beginning of the interstadial (Ammann, 2000). Plants require suitable soil conditions, and prior to 15 ka BP, soils were thin and undeveloped, although a widespread change in lake sediment stratigraphies from inorganic clays to organic muds indicates that soil development and inwash occurred soon thereafter. The initial beneficiaries of warmer conditions were open-habitat species of low competitive ability, such as grasses and sedges. Evidence from absolute pollen counts of a sharp rise of pollen influx (Pennington and Bonny, 1970) indicates that this open vegetation responded by increasing its density and biomass. Changes in species composition occurred soon afterwards, as secondary colonisers in the plant succession invaded. These were, for the most part, shrubs such as juniper (*Juniperus*), willow (*Salix*) and crowberry (*Empetrum*). Open or scrub vegetation was to dominate in Ireland, Scotland and northeast England for the remainder of the interstadial (Huntley and Birks, 1983), but in most of England and Wales, it was succeeded towards 14 ka BP by birch woodland, with downy birch (*Betula pubescens*) being the most important species. Some sites suggest a brief cold snap comparable to the Danish–Scandinavian Older Dryas around 14 000 years ago, but the consequences of this mini-stadial for vegetation

were short lived and localized.

Table 3.4 Late-Glacial stratigraphy in the North Atlantic region

Age (ka BP)	Greenland	Britain	Northern Europe	Climate
After 11.7	Holocene	Holocene	Holocene	Warm
11.7–12.9	Greenland stadial 1	Loch Lomond stadial	Younger Dryas stadial	Cold, glacial re-advance
12.9–14.7	Greenland interstadial 1	Windermere interstadial	Bölling–Alleröd interstadial	Warm
14.7–17.5	Heinrich 1 event	Devensian glaciation	Oldest Dryas	Cold
Before 17.5	Last Glacial Maximum	Devensian glaciation	Weichselian glaciation	Glacial

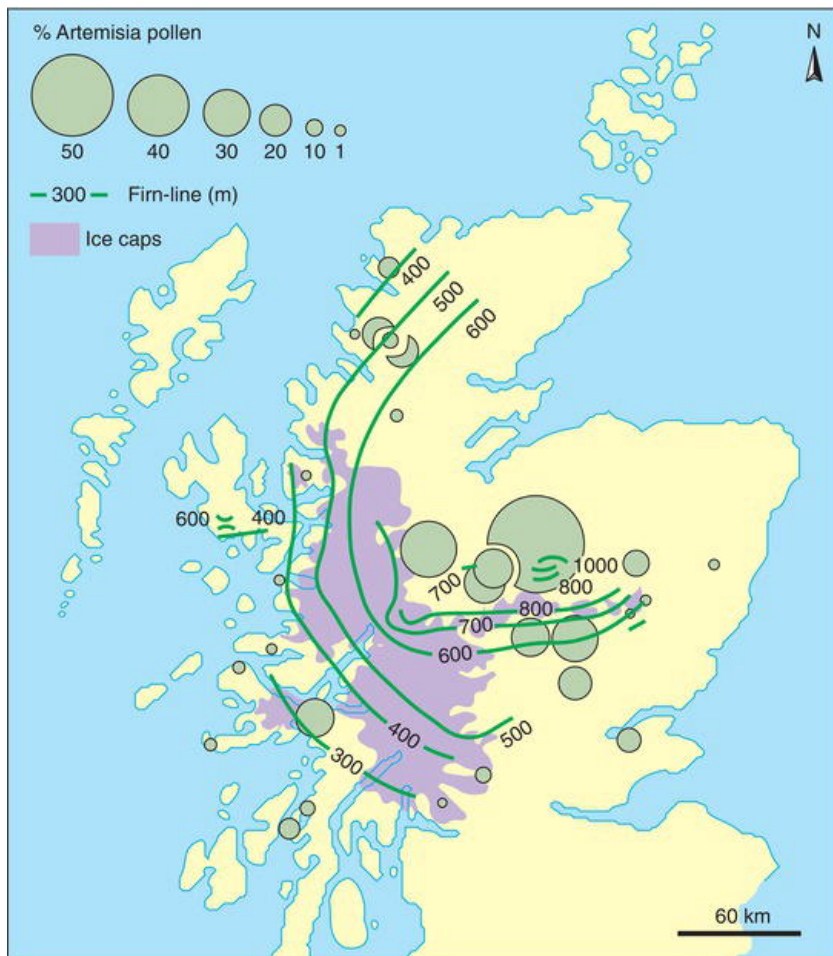
A much more significant climatic reversal began around 13 ka BP, as temperatures declined and arctic conditions returned across northwest Europe. Over Scotland, which was probably completely deglaciated during the Windermere interstadial, snow and ice accumulated once again, and by c.12.7 ka BP formed an ice cap several hundred metres thick over the western Highlands. This period, named the Younger Dryas (or, in Britain, the Loch Lomond) stadial, produced some of the best preserved glacial and pro-glacial landforms to be found in the British Isles. They include not only moraine ridges and eroded cirques but also the famous ‘parallel roads’ of Glen Roy. The latter marked the shorelines of a former ice-dammed lake whose water levels were controlled by the height of its point of outflow. Smaller valley or cirque glaciers formed in other upland areas, while at lower elevations, periglacial processes were active. Permafrost is indicated from fossil pingos and ice-wedge polygons, and these ground conditions rapidly destroyed the partially developed interstadial soils. Lake sedimentation once more became dominantly minerogenic (see [Plate 3.2](#)).

The return of a glacial climate not surprisingly threw vegetation successions into reverse. In Britain, birch woodland became restricted to locally favourable habitats, to be replaced by open-habitat taxa able to survive on disturbed and seasonally frozen soils, such as chenopods (goosefoot family). Vegetation patterns became spatially differentiated on the basis of aspect, altitude, climatic gradients and other factors (Pennington, 1977). For instance, the proportion of *Artemisia* (wormwood/mugwort) pollen at Scottish sites during the Younger Dryas stadial has been shown to co-vary with the height of the reconstructed firn line (see [Figure 3.5](#)). This has plausibly been explained in terms of a marked precipitation gradient across the country, with *Artemisia* pollen most abundant in the dry, snow-shadow zone to the north and east (Birks and Mathewes, 1978; Sissons, 1979). Beetle faunas confirm pollen and geomorphic evidence that much of northwest Europe was a tundra-like landscape in the Younger Dryas and that this was a product of a cold and generally dry climate.

Plate 3.2 A 6 m core from Lake Windermere, England, laid out in 1 m long sections. The light-coloured minerogenic sediment in the nearest section was deposited immediately after the ice retreated (c.16 ka BP); the darker brown, organic muds which follow this belong to the subsequent interstadial (14.5–13 ka BP), while the pink minerogenic sediments in the middle of the third section mark a return to a cold climate in the Younger Dryas stadial (12.9–11.7 ka BP); the topmost three sections of dark brown, organic mud cover the Holocene proper.



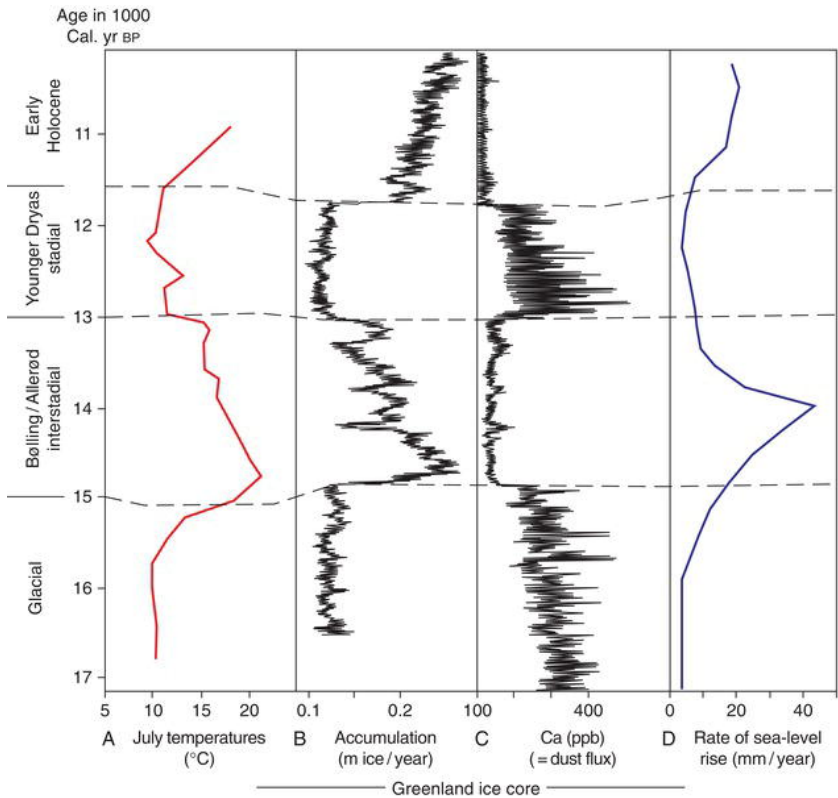
Figure 3.5 The Loch Lomond (Younger Dryas) stadial in Scotland (data from Sissons, 1979, and Tipping, 1985). Coastlines are modern.



A very similar sequence of deglacial climatic change is evident throughout the North Atlantic region, including in the Greenland Summit ice cores (Figure 3.6). These show that ice accumulated faster during warm periods, because a warmer atmosphere can hold more moisture, but the greater snowfall was more than offset by increased loss of ice through ablation and calving. The ice cores confirm the fossil invertebrate evidence in showing that some of the climatic changes during the last glacial-to-interglacial transition were startlingly rapid, including a warming of 7°C in only 50 years at the end of the Younger Dryas. From the North Atlantic Ocean itself, deep-sea sediment cores show that the polar front shifted as far north as Iceland during the warmer interstadial (15–13 ka BP) only to descend southwards again to Iberia in the subsequent stadial (13–11.7 ka BP) (Ruddiman and McIntyre, 1981). The same pattern of change also affected the contiguous land areas of western Europe (e.g. Turner and Hannon, 1988; Lowe et al., 1994; Allen et al., 1996), the Alps (Lotter et al., 1992; Lane et al., 2012) and the Atlantic seaboard of Canada and New England (LaSalle and Shilts, 1993; Mayle

and Cwynar, 1995). Pollen diagrams from the Mediterranean Basin similarly indicate a reversal in the re-advance of woodland vegetation at the end of the Pleistocene (Rossignol-Strick, 1995; Watts et al., 1996). Further eastwards and southwards, however, the climatic signal is sometimes less pronounced than in northwest Europe (Coope and Lemdahl, 1995), suggesting that across Europe the thermal oscillation varied regionally in its intensity.

Figure 3.6 The Late-Glacial climatic oscillation: **A** mean July temperatures for Britain based on Coleoptera (data from Atkinson et al., 1987; Lowe et al., 1995; note that the true rate of warming at the start of the Holocene was faster than shown here); **B** ice accumulation rate; and **C** atmospheric dust, in the GISP2 summit ice core from Greenland (adapted from Alley et al., 1993; Mayewski et al., 1994); **D** ice sheet melting as measured by the rate of sea-level rise in Barbados corals (data from Fairbanks, 1989; Bard et al., 1996).



Terminal Pleistocene climatic oscillation: A globally synchronous event?

It is natural that having been recognised in one region, changes such as the Late-Glacial climatic oscillation will be sought and found elsewhere in the world (cf. Peteet, 1993). There is danger of a reinforcement syndrome here, so it is

important that evidence from other areas is considered carefully and on its own merits rather than on the goodness of fit with the climatic sequence from the North Atlantic.

In many parts of the tropics, the period 15–11.7 ka BP was one of marked climatic instability (Gasse et al., 1990; Kudrass et al., 1991; Islebe et al., 1995; Thompson et al., 1995; Hughen et al., 1996; Hodell et al., 2008). In the northern Andes, high-altitude pollen diagrams show an irregular transition from Páramo grassland to Andean forest (see [Figure 3.4](#)), related primarily to fluctuations in temperature. However, moisture variations may have been as important as temperatures in controlling the water balance of tropical and subtropical lakes. In Africa, lake-level fluctuations represent some of the strongest evidence for terminal Pleistocene climatic instability. This was a transitional period in the lake record, many lake levels rising after 15 ka BP only to fall again at around 13 ka BP (Street-Perrott and Roberts, 1983; Roberts et al., 1993; Hoelzmann et al., 2004). Low water levels in the tropics reflect an arid phase which was a product of the same disturbance to the climatic system that produced the northwest European Younger Dryas stadial (13–11.7 ka BP). The hydrological cycle was weakened at this time, probably because glacial meltwater reduced evaporation from oceanic surface waters in the northern hemisphere. In western North America too, lake-level and pollen data imply that cool dry events punctuated the longer-term glacial to interglacial climatic transition (Benson et al., 1997).

One of the most important pieces of evidence for the Younger Dryas having global consequences comes from the sea-level record. Richard Fairbanks (1989) carried out detailed dating of cores taken through drowned coral reefs in Barbados, which normally grow within a few metres of the ocean surface. From this he built up a history of sea-level rise since the LGM, which showed a significant slowdown between 13 and 11.7 ka BP (see [Figure 3.6](#)). In a region like the Caribbean, sea levels were primarily controlled by eustatic factors and therefore closely reflect the build-up and decay of ice sheets. However, a slowdown in the rate of global sea-level rise during the Younger Dryas does not in itself tell us which of the world's former major ice sheets had stopped melting. This, in turn, raises the wider question of whether deglaciation was in or out of phase between the planet's two poles.

Evidence from Antarctica suggests it was not; in other words, climate change during the last deglacial transition was not synchronous between the two hemispheres. In ice cores from both Greenland and Antarctica, a short-lived cold reversal interrupted the overall warming trend after the LGM, but not – it appears – at the same time. The Antarctic Cold Reversal is dated to around 14–13 ka BP, at the same time as the North Atlantic was experiencing interstadial warmth! If correct, then during the last great global warming, the southern hemisphere led the northern hemisphere by a thousand years or more. Reliable dating and correlation are critical to this assessment, and global chronologies are notably less certain during the Late-Glacial period than they are in the Holocene. Even within the North Atlantic region, and using tephra layers from volcanoes in Iceland, Germany and elsewhere to act as regional marker horizons, it has proved difficult to establish precisely agreed dates for key events such as the abrupt warming at the start of the Bølling interstadial (Björck et al., 1998). Nonetheless, there are good grounds for believing that sub-millennial climatic changes during the late

Pleistocene were out of phase between Antarctica and Greenland. One especially persuasive piece of evidence comes from the methane record in polar ice cores. Methane, which under natural circumstances came mainly from sources on land in the tropics or northern hemisphere, is mixed rapidly through the global atmosphere, so records of its changing atmospheric concentration must have had the same trends at the same time everywhere on the planet. In ice cores from Antarctica, methane concentrations reached a pre-Holocene peak at the same time as temperatures were cooling (Monnin et al., 2001), whereas in the GRIP ice core from Greenland, the same methane peak matches the period of maximum Late-Glacial warmth (Blunier et al., 1995). Whatever its precise age, this methane peak must be the same event in both sequences, and it demonstrates that a temperature see-saw effect operated between the two polar regions during the transition from a glacial to an interglacial Earth (Blunier and Brook, 2001). As [Figure 3.3](#) illustrates, this climatic see-saw between the planet's poles seems to have operated throughout glacial times, with D-O warm-climate interstadial events in Greenland coinciding with cold stadial conditions in Antarctica. Glacier histories and pollen evidence from areas such as New Zealand and southernmost South America seem to be in tune with the record from Antarctica (Clapperton, 1993; Heusser, 1993; Turney et al., 2003; Hajdas et al., 2006; Denton et al., 2010; García et al., 2012; Newnham et al., 2012), whereas most of the northern hemisphere and the tropics changed in synchrony with Greenland and the North Atlantic.

What caused this erratic and complex pattern of global change? Ice core records from Greenland show that northern hemisphere temperatures changed in close harmony with atmospheric greenhouse-gas concentrations, providing evidence for a link between the two. The same correlation is evident in stomatal densities in plant leaves dating from the Younger Dryas, which are believed to be the result of atmospheric CO₂ levels (Beerling et al., 1995). However, because this relationship did not hold true in Antarctica, greenhouse gases are unlikely to have been the main global cause of the rapid 'flip-flops' in climate during the last glacial to interglacial transition. Instead, it is more likely that the sudden jumps in climate were linked to changes in the world's ocean circulation and in particular to part of the 'conveyor belt' circulation in the Atlantic Ocean periodically switching on and off (Broecker et al., 1985). The Atlantic ocean conveyor carries warm surface water northwards which afterwards returns southwards as cold, dense deepwater, keeping Europe several degrees warmer than equivalent latitudes elsewhere. The present operation of the conveyor is delicately balanced depending on the salt budget of the Atlantic Ocean. A small decrease in the salinity of North Atlantic surface waters would be enough to prevent it sinking to form deepwater, the conveyor would then stop and Europe would suddenly become much colder. Because this thermohaline circulation feeds into the rest of the world's ocean circulation, switching it on or off would also affect climates well beyond the North Atlantic (Street-Perrott and Perrott, 1990). Wally Broecker (1994) proposed that large pulses of freshwater from melting North American and European ice sheets would have changed the salt budget of the Atlantic sufficiently for the 'conveyor belt' circulation to be temporarily switched off. The Younger Dryas climate reversal, which bucks the longer-term Milankovitch trend, was most probably the result of a shutdown or weakening of the Atlantic

conveyor coinciding with longer-term northern hemisphere deglaciation. A switch to the Atlantic conveyor may have been triggered by the diversion of meltwater from the Mississippi (and thence into the Gulf of Mexico) to a spillway eastwards via the St Lawrence into the North Atlantic (Teller et al., 2002) or via the McKenzie river into the Arctic Ocean (Murton et al., 2010). By contrast, the circulation of the Southern Oceans is influenced most strongly by the formation of deepwater from Antarctic, and which is coupled to the Atlantic conveyor via a bipolar see-saw mechanism (Broecker, 1998). This mechanism caused heat and salt to be moved alternately north and south throughout much of the last glacial stage, before eventually stabilizing once deglaciation was complete. Ice and ocean therefore combined to respond in complex fashion to a gradual forcing of the climate, resulting in step jumps between different states of the Earth system (Alley and Clark, 1999; Alley, 2007). One consequence was that the last glacial period ended, not gradually, but with extreme abruptness – probably within a human lifetime – as the Earth system crossed a tipping point between a glacial state and an interglacial one (Steffensen et al., 2008).

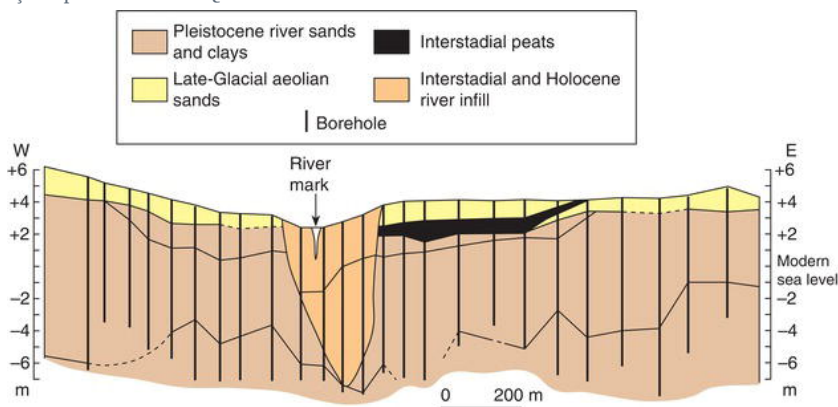
One further element that has been added to the enigma of the Younger Dryas climate reversal is the possibility that it was triggered by the impact over North America of a comet or other object(s) from outer space around 13 ka ago (Firestone et al., 2007). This could have acted to rapidly cool the climate and may also have had consequences for megafauna (see following text). There are echoes here of the dinosaur extinction event at the end of the Cretaceous period, 65 million years ago, which has been linked to the collision of an extraterrestrial body with the Earth. Unlike at the Cretaceous–Palaeogene (K-T) boundary, however, no crater has been found associated with the – hypothetical – Younger Dryas impact event. The evidence instead comes from micro-charcoal evidence of extensive wildfires, along with unusually high concentrations of materials that are normally rare on Earth, such as so-called nano-diamonds (the product of very high temperatures) and chemical elements such as iridium. A Late-Glacial extraterrestrial impact is not an impossibility; an air burst of a large meteoroid or comet fragment occurred 5–10 km above Tunguska in Siberia in 1908 flattening an estimated 80 million trees! However, the proposed pre-Younger Dryas impact event is likely to remain controversial until or unless it is confirmed by unambiguous evidence.

Adjustment of geomorphic systems

Because soil formation and vegetation lagged behind the often rapid shifts in climate, many landscapes experienced a phase of temporary geomorphic instability during the terminal Pleistocene. In addition to the many landforms created directly by deglaciation, such as eskers and kettle holes, there were also those produced indirectly, for example, by glacial meltwater. For example, a spectacular pro-glacial drainage system existed in the Columbia river system in the northwest United States, where repeated draining of pro-glacial Lake Missoula caused short-lived floods with discharges up to 20 times total modern global runoff – the largest freshwater flows so far known on Earth (Baker and Bunker, 1985; O'Connor and Baker, 1992; Smith, 1993)!

Braided stream deposits reflect increased discharges of both water and sediment that were widespread at the end of the last glaciation (Rose et al., 1980; Baker, 1983; Maizels and Aitken, 1991). These palaeohydrological changes often led to dramatic alterations in river morphology, a process termed metamorphosis by Stanley Schumm (1969). In alluvial plains, such as those of the Murray Basin in Australia, large channels containing sandy sediments were replaced by sinuous silty clay ones at the start of the Holocene (Page et al., 2009). Many streams and rivers which experienced increased discharges during the terminal Pleistocene are now underfit. George Dury calculated that ratios between modern and former meander wavelengths in North America vary from 1 : 10 near the ice front in Wisconsin to 1 : 3 in the more southerly Ozark Mountains. These, he proposed, reflect late Pleistocene discharges up to hundred times greater than at present (Dury, 1965). At a smaller scale, the dry valleys of southern England are also relict features from the Pleistocene, in this case, having been produced by solifluction and related processes during seasonal snowmelt (Kerney et al., 1963).

Figure 3.7 Cross section through an alluvial valley fill, R. Mark, Belgium/Netherlands (after Vandenberghe and Bohncke, 1985). Reproduced with permission of Bulletin de l'Association française pour l'étude du Quaternaire.



The unvegetated glacial outwash plains of the terminal Pleistocene were vulnerable to deflation by the wind, and aeolian sand and loess were actively transported and widely deposited as the ice sheets retreated (Catt, 1978; Wells, 1983). The small river valleys of the Dutch–Belgian border, for instance, became choked with wind-blown sand at this stage (Vandenberghe and Bohncke, 1985). After 15 ka BP, the cold, dry climate and near absence of vegetation gave way to the warmer conditions of the Late-Glacial interstadial, and river palaeohydrology changed strikingly (Vandenberghe, 2008). In the case of the River Mark valley shown in Figure 3.7, sediment load diminished as aeolian activity ceased and slopes became stabilized by soil and vegetation, while at the same time river discharges increased. In combination, these resulted in sharp incision of the pre-existing valley fill to depths of over 8 m. Although these incised meanders soon began to fill up with sediment, the fluvial system remained in a disturbed state with localised blocking of the valley by sand, ponding back of rivers and peat formation. Only around 10–9 ka BP did the geomorphic disequilibrium of the

Pleistocene–Holocene transition give way to a regular longitudinal river profile (Vandenberghe and Bohncke, 1985). In general, Holocene fluvial systems in temperate latitudes have been characterised by lower-energy environments that had existed previously and by a sediment supply that came to be regulated by the well-developed forest and grassland cover (Knox, 2000). In consequence, Holocene alluvial fills in temperate latitudes are typically fine grained in nature (Gregory et al., 1987; Williams, 2012).

Intense geomorphic activity during the Pleistocene–Holocene transition also characterised many low-latitude regions. Although the climatic transition here was from arid to humid rather than from periglacial to temperate, it created the same kind of disequilibrium between vegetation and climate. Unvegetated, erodible soils combined with high rainfall erosivity to produce the highest sediment yields of the whole of the late Quaternary prior to recent human impact. These yields are recorded by increased rates of sedimentation in coastal deltas, inland lakes and river terrace sequences (Roberts and Barker, 1993; Thomas and Thorp, 1995). The Nile, which had been a highly seasonal, braided river between 23 and 15 ka BP, experienced a huge increase in discharge, partly on account of overflowing East African lakes expanding the river's catchment area (Williams et al., 2010).

Human ecology at the end of the Pleistocene

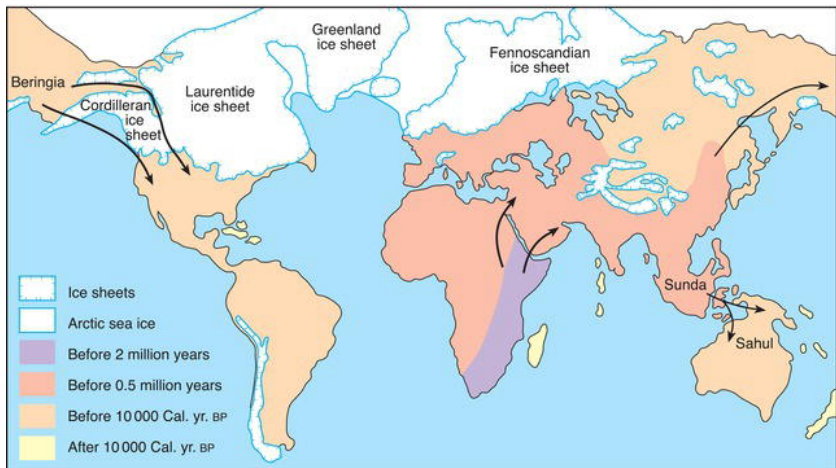
The probable birthplace of the human race lies in East Africa, around five million years ago. From here the genus *Homo* spread into all other continents save Antarctica by the start of the Holocene. However, establishing the precise date at which **hominin** populations first arrived at a particular part of the Earth's surface is problematic, not least because new discoveries are constantly pushing these dates further back in time. It is perhaps safer to consider the latest agreed rather than the earliest possible dates of arrival, and using this approach a three-stage model for human expansion can be suggested. For the first few million years, hominins were restricted to eastern and southern parts of the African continent, but by half a million years ago, they had spread into the adjacent continents of southern Asia and Europe (Gamble, 1993; Stringer, 2002).

The second stage in the expansion came significantly later and involved the movement of anatomically modern humans out of Africa into Eurasia, followed by the first peopling of Australia and the Americas (Oppenheimer, 2004). In the latter continents, there were formidable natural barriers to be overcome which delayed human colonisation until late in the Pleistocene. To reach Australia–New Guinea meant a sea crossing by raft or canoe of at least 100 km even at times of low sea level. Remarkably, this was achieved around 40–50 000 years ago as indicated by ¹⁴C- and luminescence-dated burial sites such as Lake Mungo (Jones, 1989; Gillespie, 2002; Bowler et al., 2003; O'Connell and Allen, 2004). By 30 ka

BP, all of Australia's main ecological zones had been occupied, along with the adjacent islands. Palaeo-Indians arrived somewhat later in the Americas than the Aboriginal populations did in Australia. It was possible to walk from Asia to America via the dry land bridge of Beringia (Elias et al., 1996), but once in Alaska, any potential colonisers would have found their way blocked by the giant North American ice sheets (see [Figure 3.8](#)). Nor would further progress have been encouraged by the presence of some ferocious Ice Age predators roaming the Yukon and Alaskan tundra, including a gigantic carnivorous bear (*Arctodus simus*) (Kurtén and Anderson, 1980; Geist, 1989, 1999). By the time the route south opened up, rising sea levels had cut the way off back to Asia. It is therefore not entirely surprising that the main expansion of human population in the Americas did not take place until the terminal Pleistocene around 13 000 years ago (Waters and Stafford, 2007). On the other hand, there is an active debate as to whether these hunters of the Clovis and other late Pleistocene cultures were home grown from pre-existing, pioneering groups, for which there is some fragmentary archaeological evidence, or whether they were of exogenous origin, descending from Alaska via an ice-free corridor that opened up during deglaciation or via coastal island chains (Josenhans et al., 1997; Haynes, 2002; Meltzer, 2009).

The third and final stage in the colonisation of the Earth took place during the Holocene and involved the inhospitable terrain around the ice sheets of Greenland and northern Canada, and many of the world's islands. It is curious that despite the existence of waterborne craft, many of the Mediterranean islands were not settled until after the end of the last glaciation (Cherry, 1981; Schule, 1993). More easily understood is the late arrival of humans in the geographically isolated islands of the Pacific (from c.3500 Cal. yr BP) and Madagascar (c.2000 Cal. yr BP; Burney et al., 2004).

Figure 3.8 The human colonization of Ice Age Earth. Coastlines shown are for glacial maxima. Arrows show key pathways for human dispersals.



The Pleistocene colonisation of the Earth by humans was an extraordinary achievement (Cavalli-Sforza and Cavalli-Sforza, 1995). What is more, it was

achieved with subsistence economies entirely based on hunting of wild animals, fishing and gathering of wild plants. Hunter-fisher-gatherer (h-f-g) peoples need to have an intimate knowledge of their natural environment upon which they depend for food and other resources. 'Man the hunter' must be familiar with the movement and behaviour of his quarry, whether stag or seal, just as 'woman the gatherer' must know of the location and uses of different plants, whether hemlock or hazelnuts. Ethnographic studies of modern h-f-g groups suggest that plant foods are normally a more important component of the diet than meat, although both this and a gender-based division of labour may not have been true of all Pleistocene populations. Establishing the relative importance of the hunted/scavenged versus the gathered food component at archaeological sites is not easy, because plant remains are much less well preserved than animal bones. However, it is likely that meat provided a large share of the food supply to later Pleistocene populations in mid-latitude Eurasia. This is because the open steppe-tundra glacial vegetation would have supported large numbers of game but provided little in the way of edible plants. Hunting groups were able to follow animal herds across the open landscape, relying on them to provide protein, fat, hide, bone and all the other material necessities for life. By contrast, the forest environments of the tropics and of the Holocene possessed a wide range of potential food plants which may have shifted the dietary emphasis away from meat. H-f-g bands move camp throughout the year following game, water or other resources, and need a relatively large territory to support them, at least 75 km² for a band of only 25 persons. The resulting low population densities mean that even at the end of the Pleistocene, world population is unlikely to have risen above ten million. Because this level was maintained for many thousands of years, however, 90% of all humans ever to have been alive would have been hunters, fishers or gatherers.

Simple Stone Age technology and low population numbers limited the extent of human impact upon the natural environment. Plant resources were manipulated by agencies such as fire, and burning to clear vegetation or encourage regrowth is certainly one of our oldest tools of environmental management (Lewis, 1972; Woodcock and Wells, 1994). Although forest or grassland fires only take hold where there is abundant dry plant matter, the spark to light them could be caused by lightning strikes as well as by rubbing fire sticks. It is therefore difficult to know how far former fire frequencies were naturally or culturally determined. Significantly, in a long pollen record from northern Australia, charcoal frequencies are many times higher during the last 40 ka than they were during earlier parts of the late Pleistocene, when Australia's aboriginal population had not yet arrived (Kershaw, 1986; Rule et al., 2012). However, more clearly evident than the effect on vegetation was human impact on other parts of ecosystems, notably upon large mammals who were the prey or competitors of *H. sapiens*.

Megafaunal extinctions

Towards the end of the Pleistocene, there occurred a devastating wave of animal extinctions (Elias and Schreve, 2007). The most obvious victims were mammals with body weights over 44 kg, such as the mastodon (*Mammuth americanum*), the

woolly rhino (*Coelodonta antiquitatis*), the giant deer (*Megaloceros*) and the native American horse (*Equus occidentalis*) (see [Plate 3.3](#)), but birds and smaller mammals were not entirely unaffected either (Grayson, 1977). The extinctions varied in their timing and severity across the globe, but the most widespread phase took place at the end of the Pleistocene (16–11.7 ka BP) and affected the Americas and northern Eurasia (see [Table 3.5](#)). As we have seen, this was a period of rapid climatic change, and it has been suggested that many large mammals struggled to adapt to the new environmental conditions (Reed, 1970). For example, a family of mammoths (*Mammuthus primigenius*) whose remains were found at Condover in England have been ^{14}C dated to precisely the period of most rapid warming just after 15 ka BP (Coope and Lister, 1987). Herbivores such as these found their tundra-steppe environment being displaced northwards and reduced in extent by the invasion of woodland. In turn, carnivores like the sabre-toothed tiger (*Smilodon*), who preyed upon grazing mammals, found their food resources diminishing. In some cases, movement to new areas was hampered by mountain ranges, rising sea levels or other natural barriers. This combination of climatically caused environmental stresses created population bottlenecks which could have led to the more specialised and less adaptable species being exterminated.

Plate 3.3 Skeleton of a woolly mammoth, one of many species of megafauna made extinct during the late Pleistocene. This specimen, found on the banks of the river Shandra in eastern Siberia, is one of the largest complete skeletons found so far (the distance between the tusk ends is 3 m!). In addition to the skeleton, this mammoth's heart, liver, stomach and intestines were also preserved as a frozen mass weighing 700 kg. Photo: A. Zubtsov. Reproduced with permission of Novosti Press Agency.

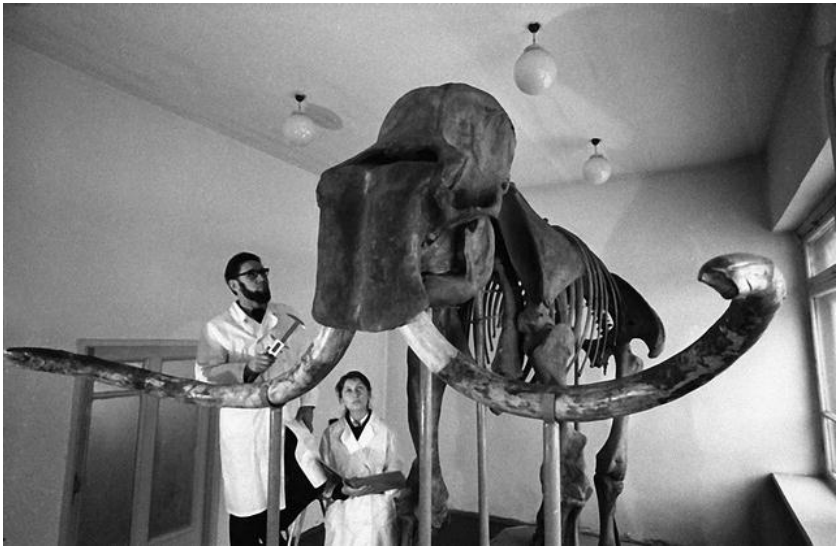


Table 3.5 Late Pleistocene extinct and living genera of large terrestrial mammals (> 44 kg adult body weight)

	Extinct	Living ^a	Total	Extinct (%)	Major period of extinction (ka BP)
Australia	13	3	16	81	50–40
South America	46	12	58	80	15–9
North America	33	12	45	73	16–11.5
Europe ^b	9	14	23	39	16–10
Africa	7	42	49	14	14–10.5

^aOr extinct in historical times
^bExcluding Mediterranean islands
Source: Adapted from Martin and Klein (1984)

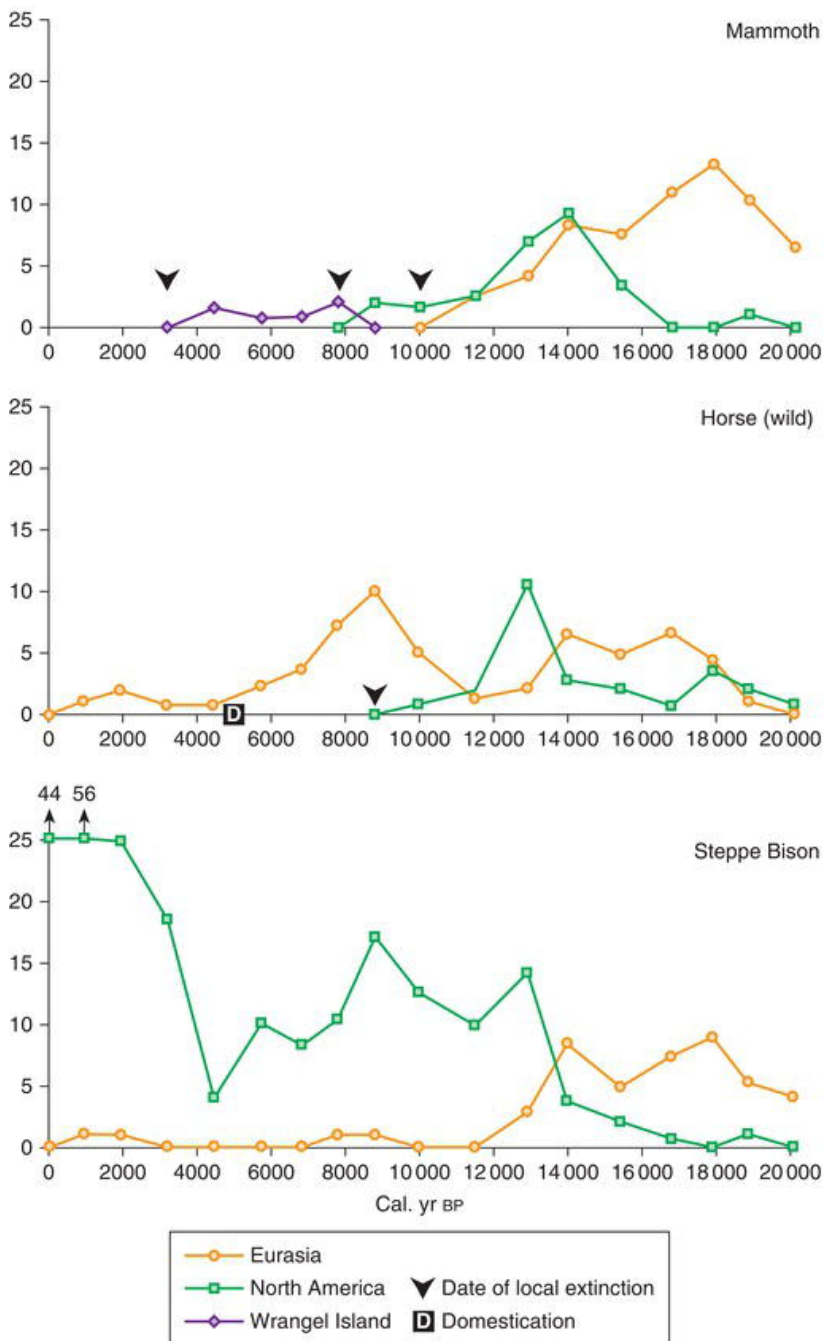
Unfortunately, on its own, this hypothesis has some manifold weaknesses. Most obvious of these is the fact that no comparable extinctions took place at other times of rapid climatic change during the Pleistocene, for instance, at the beginning of the last interglacial c.130 000 years ago. Researchers such as Paul Martin have instead been struck by the coincidence between changes in human culture and the timing of megafaunal extinctions. This is most clearly the case in North America where the appearance of the **Clovis culture** at around 13 ka BP was followed within less than a thousand years by the demise of many large mammals. It has been argued that big game hunters of the terminal Pleistocene butchered, in vast numbers, animals that were quite unused to human predation (Martin and Klein, 1984). Archaeological support for this overkill hypothesis comes from ‘kill’ sites at which a great many animal bones are present, although perversely such associations are much more common in the Old World than the New. For instance, the bones of over 100 000 horses (*Equus przewalskii*) have been found at the Upper Palaeolithic butchery site at Solutré in France. Later Holocene island extinctions, such as that of the giant flightless moa bird (*Dinornis*) (Plate 7.4), similarly followed within a few centuries of the arrival of humans, in this case, the Maori peoples in New Zealand (Worthy and Holdaway, 2003). Evolution in isolation meant that prior to the Holocene, the Mediterranean islands were home to creatures almost as bizarre as those found in Dr Dolittle’s zoo, including flightless swans, dwarf ‘antelopes’ (*Myotragus*) and pygmy hippos. With the arrival of the first permanent human populations in Neolithic times, all of these rapidly became extinct (Lewthwaite, 1989; Simmons, 1991; Schule, 1993). Not a single one of the larger ‘wild’ animals on islands such as Mallorca and Cyprus is therefore truly native; all have been introduced by people.

Although there are good grounds for believing that human agency was involved in most cases of faunal extinction, overkill seems no more satisfactory a universal explanation than does climatic change. Extinction of the mammoth and the mastodon may have followed soon after the appearance of the Clovis hunters in America, but in Europe and Asia, hunting populations coexisted with megafauna for many thousands of years before the latter declined at the end of the Pleistocene. With animals such as the giant deer (*Megaloceros*), which was abundant in Ireland during the Late-Glacial period, extinction seems to have occurred before humans had even arrived (Stuart et al., 2004). Another feature

inconsistent with overkill is that many of the animals which are most numerous at archaeological sites managed to avoid extinction, for instance, the bison (*Bison bison*) in North America and the reindeer (*Rangifer tarandus*) in Europe. These lines of evidence suggest that indirect human action may have been as important as direct kill-off (Krantz, 1970). As hunters, late Pleistocene humans would have partly replaced the ecological niche of carnivores and may also have threatened non-favoured herbivores such as the American Shasta ground sloth (*Nothrotheriops shastense*). In contrast, favoured game species would have been preserved, probably by laws and taboos similar to those recorded historically for hunter-gatherer groups. This would have applied especially to herd ungulates like the bison and the reindeer with which human populations were able to develop a symbiotic relationship (see [Figure 3.9](#)).

Having modified the structure and stability of ecosystems, human populations would have made large mammals more vulnerable to other environmental stresses. In particular, the changing and, from the faunal point of view, deteriorating climate towards the end of the last glaciation would have created population 'bottlenecks'. Whereas creatures such as the mammoth had survived through previous climatic bottlenecks, on this occasion, animals were forced to compete with humans for increasingly scarce resources. In western North America, for example, a contraction in water resources occurred at around the time of the Younger Dryas, and Clovis culture artefacts and mammoth tracks have been found together around spring sites (Haynes, 1991). The 'double whammy' of human competition and climatic stress, possibly along with other factors like disease, seems to have proved terminally fatal for many animal species (Barnosky et al., 2004; Surovell et al., 2005). Indirect human disturbance of ecological balances is also consistent with the differing severities of late Pleistocene extinctions across the world. In Africa, where hominins had formed part of savanna ecosystems for millions of years, a phase of extinctions can hardly be recognised at all. Nearly all of that continent's megafauna has remained intact until the present day, to fill the game parks and cages of modern zoos. In northern Eurasia, some eight genera disappeared at the end of the Pleistocene, all of them adapted to steppe-tundra environments and most of them exploited by Stone Age hunters through the last glaciation (Stuart, 1999). In the Americas, where human antiquity is shorter and ecosystems were probably pristine until the terminal Pleistocene, mammalian extinctions were devastating in their effects. They involved 39 genera of large mammals in North America alone, a greater number of extinctions than during the preceding four million years (Martin and Klein, 1984), and they have left New World megafaunas impoverished ever since.

Figure 3.9 ¹⁴C-dated occurrences of three megafaunal taxa in the New and Old Worlds (data from McDonald, 1984; Dolukhanov and Khotinskiy, 1984; Vereshschagin and Baryshnikov, 1984; Vartanyan et al., 1993).



Extinctions were equally dramatic in Australia. Here dating evidence has gradually pushed their timing back from the end of the Pleistocene to around 40–

50 000 years ago (Miller et al., 1999; Roberts et al., 2001). As this coincides with the first aboriginal arrivals on that continent, it strongly supports a link between humans and the decline and demise of Australian megafauna (Flannery, 1994). Miller et al. (2005) have suggested a mechanism for how this might have occurred in the dry, fire-prone Australian outback. They obtained a 140 000-year record of emu diet from $\delta^{13}\text{C}$ analysis of preserved eggshells which showed a permanent reduction in this flightless bird's food sources from the time of human colonisation of Australia. Combined with pollen and charcoal evidence for increased burning of the vegetation after human colonisation, Miller et al. (2005) speculate that the pre-existing mosaic of trees, shrubs and nutritious grasslands was rapidly converted to the modern fire-adapted desert scrub. At the same time, a sharp drop-off in the spores of a fungus that lives in animal dung indicates that large herbivores declined rapidly in number (Rule et al., 2012). It seems that those animals which could adapt or hide survived, while many others that could not became extinct.

Perhaps most remarkable megafaunal discovery have been the remains of dwarf mammoths of Holocene Age on Wrangel Island in the Siberian Arctic (Vartanyan et al., 1993) and on St Paul Island in the Bering Strait (Guthrie, 2004). During the late Pleistocene, these islands were joined to the mainland by land bridges, but as sea levels subsequently rose, they became isolated, and their mammoth populations were trapped. Everywhere else in the world, mammoths quickly became extinct, but on these isolated islands, mammoths were able to survive in an Ice Age time capsule of tundra meadows and arctic steppe, untroubled by natural predators or by humans. As often happens with isolated populations the mammoths started to 'shrink' in size, in this case to around half their previous height. Numerous bones, tusks and teeth of these dwarf mammoths have been ^{14}C -dated to the early and mid-Holocene, but – sadly – they show that these mammoth populations eventually succumbed, probably to natural causes. The youngest dates, however, indicate that on these remote Arctic islands, mammoths survived until after 4500 years ago, by which time Egypt's great pyramids were being erected!

The human role in terminal Pleistocene extinctions remains a matter of controversy (Grayson and Meltzer, 2002; Fiedel and Haynes, 2004; Steadman et al., 2005; Koch and Barnosky, 2006). Nonetheless, it is likely that human impact upon this key part of ecosystems was substantial even before the start of the Holocene (Pinter et al., 2011). The separation of human and natural agencies therefore emerges as a problem even before climate, vegetation and landscapes became recognisably modern in form.

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Early Holocene adaptations (11 700–6000 Cal. yr bp)

Changes in the physical environment

The onset of the Holocene witnessed the start of environmental processes which have continued up to the present day; processes such as soil formation, plant succession, lake ontogeny and faunal migration. Although interglacial conditions became established soon after 11.7 ka BP, it would be quite wrong to imagine that the climates, ecosystems and landforms of the early Holocene were identical to those of the present day or that they have remained static since that time. The broad outlines of the modern natural world were established earlier in some regions and later in others, and for a time this produced strong inter-regional contrasts, sometimes even over short distances. The first set of changes to be considered will be those taking place in the physical environment during this formative stage of the Holocene.

Ice sheets and sea levels

Over northwest Europe and many other parts of the world, modern summer temperatures were established within the first thousand years of the Holocene (see [Figure 3.6](#)), and this temperature rise produced a rapid response by smaller ice caps and valley glaciers. In the Val d'Hèrens in the Swiss Alps, for example, a moraine dated to 9.4 ka BP lies adjacent to the modern Tsidjiore Nouve glacier, marking a retreat of over 20 km from the position of the glacier snout two millennia earlier (Rothlisberger and Schneebeli, 1979) (see [Plate 2.9](#)). However, because of their very bulk, the major northern hemisphere ice sheets took longer to melt away. In particular, the Laurentide ice sheet remained extensive until around 8.5 ka BP, when its heart was eaten out as the sea invaded Hudson Bay.

The retreat of the ice sheets caused [eustatic](#) sea levels to rise from about –55

m at the start of the Holocene to around modern elevations by around 6000 years ago (Fairbanks, 1989; Chappell and Polach, 1991). Up until the end of the Pleistocene, eustatic sea-level rise had accounted for most coastline changes because of the virtually instantaneous response of eustasy to ice sheet melting. Glacio-isostatic adjustment (GIA) in deglaciated regions involved a slower response, but it has played a prominent part in the Holocene history of European and North American coastal zones (see [Figure 4.1](#)). In places previously compressed under ice, the unloading of this weight caused the land literally to rebound out of the sea (see [Plate 4.1](#)). The scale of changes since the terminal Pleistocene can be judged from the fact that Late-Glacial shorelines in Scotland are now up to 40 m above sea level, representing a glacio-isostatic uplift of at least 70 m during the past 15 000 years, while in some parts of Scandinavia, Holocene uplift has been several times that amount.

Adjacent to the area of former glaciation, on the other hand, was a zone whose surface had been pushed up by the weight of the ice. This forebulge has experienced compensatory subsidence during the Holocene and forms a zone which continues to suffer submergence today. It is the increasing risk of inundation along the US eastern seaboard and the southern North Sea littoral that has led to the construction of major sea defence systems such as those that have been built in the Rhine delta region of the Netherlands. Beyond this, even far-field regions such as the tropics were not free from the effects of the redistribution of ocean water and crustal mass during and after the last deglacial transition. This, for example, accounts for the fact that sea levels in the Pacific were around 3 m higher during the mid-Holocene than they are today ([Figure 4.1](#)) (Mitrovica and Milne, 2002).

The interaction of eustatic and glacio-isostatic factors has often led to complex local histories of relative sea-level change, none more so than in the case of the Baltic (Björk, 1995). A large pro-glacial lake formed here at the end of the Pleistocene, adjacent to the retreating Fennoscandian ice mass. Towards the tail end of the last glaciation, the ice retreated from Mt. Billingen in south central Sweden and caused the Baltic ice lake to fall to what was then sea level (Björck and Digerfeldt, 1989). The Baltic became a marine embayment, named the Yoldia Sea after a brackish-water mollusc that lived in its waters. Glacio-isostatic uplift, however, outpaced the eustatic sea-level rise in this region and isolated the Baltic once more. This second freshwater body formed c.10.6 ka BP and has been named the Ancylus lake (Eronen, 1983). Finally, c.8.4 ka BP, the sea once more succeeded in flooding what are now the Straits of Denmark to produce the Littorina Sea (named after the common periwinkle), the immediate predecessor of the modern Baltic. The Baltic was thus twice a freshwater lake and twice linked to the sea over the space of five millennia.

Figure 4.1 Holocene sea-level changes: A areas of rebound and submergence (after Walcott, 1972; reproduced with permission of Elsevier); B eustatic sea level curve (data from Bard et al., 1996).

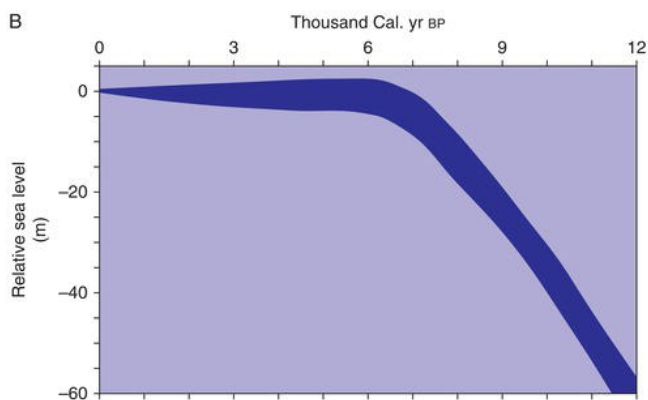
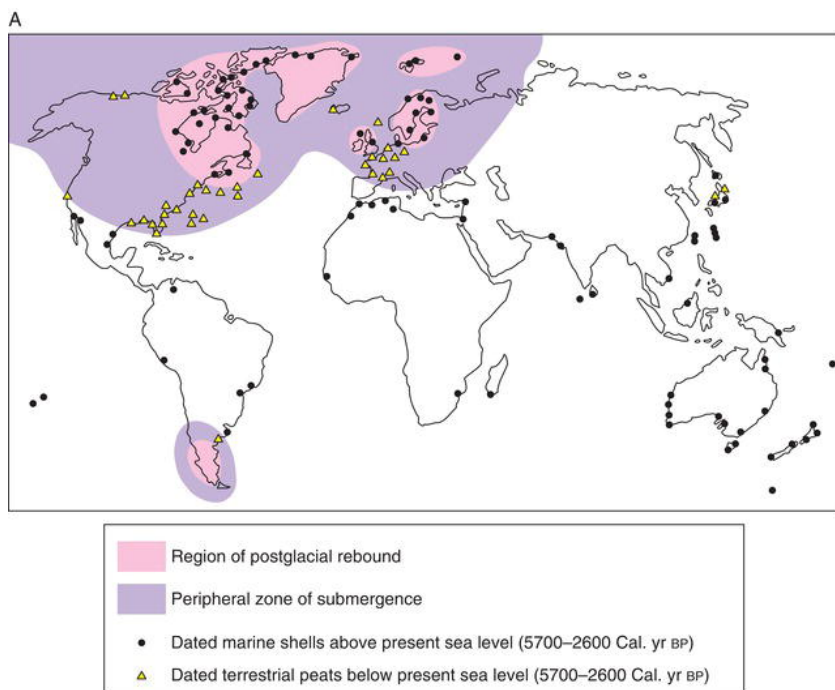


Plate 4.1 Raised marine platform, western Islay, Scotland, now 35 m above sea level after glacio-isostatic uplift. Photo by David Smith, Coventry University. Reproduced with permission of David Smith.



The configuration of the glacial coastline differed most strongly from that of today in areas with shallow offshore gradients. It was these same areas which were now most rapidly inundated by the incoming tide. The subcontinent of Sunda in southeast Asia, for example, became fragmented into the many islands of the Indonesian archipelago. In areas such as these, land was drowned at a rate that must have been noticeable from one year to the next. What is sure is that human populations had to relocate themselves and their economic activities, as archaeological sites adapted to the life of the seashore, such as shell middens, testify.

Human adaptations to coastal environments

Human utilisation of coastal resources extends well back into the Pleistocene. On the other hand, most evidence from these early periods has been drowned by Holocene high sea levels, so that it is only for the late Quaternary that we have an adequate record of coastal occupation patterns. There is evidence of maritime-based human economies in the New World as early as 12.6 ka BP from the archaeological site of Quebrada Tacahuay on the coast of southern Peru (Keefer et al., 1998). On the other side of the Pacific, in Japan, the Jomon Neolithic culture, which endured from 11.7 to as late as 2.3 ka BP, represents another adaptation to the diverse marine and littoral resources that became available as the last glaciation ended (Imamura, 1996). Around 8000–7000 years ago, the Jomon marine transgression, of tectonic as well as eustatic origin, produced sea levels 3–6 m above modern elevations (Ishiga et al., 2000). Along this former coastline, now up to 60 km inland, lie shell mounds and other Jomon occupation sites containing marine molluscs, such as *Meretrix* and *Macra*, and remains of deep-sea fish (Akazawa, 1986). Rising Holocene sea levels also served to isolate the

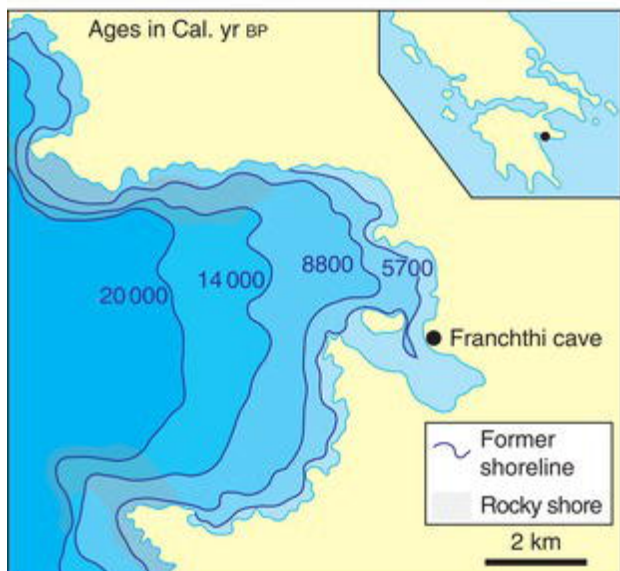
Japanese islands from the rest of East Asia. In consequence, the Jomon Neolithic, with its distinctive material culture that included pottery, is culturally quite different from any of the contemporary developments on the Asian mainland.

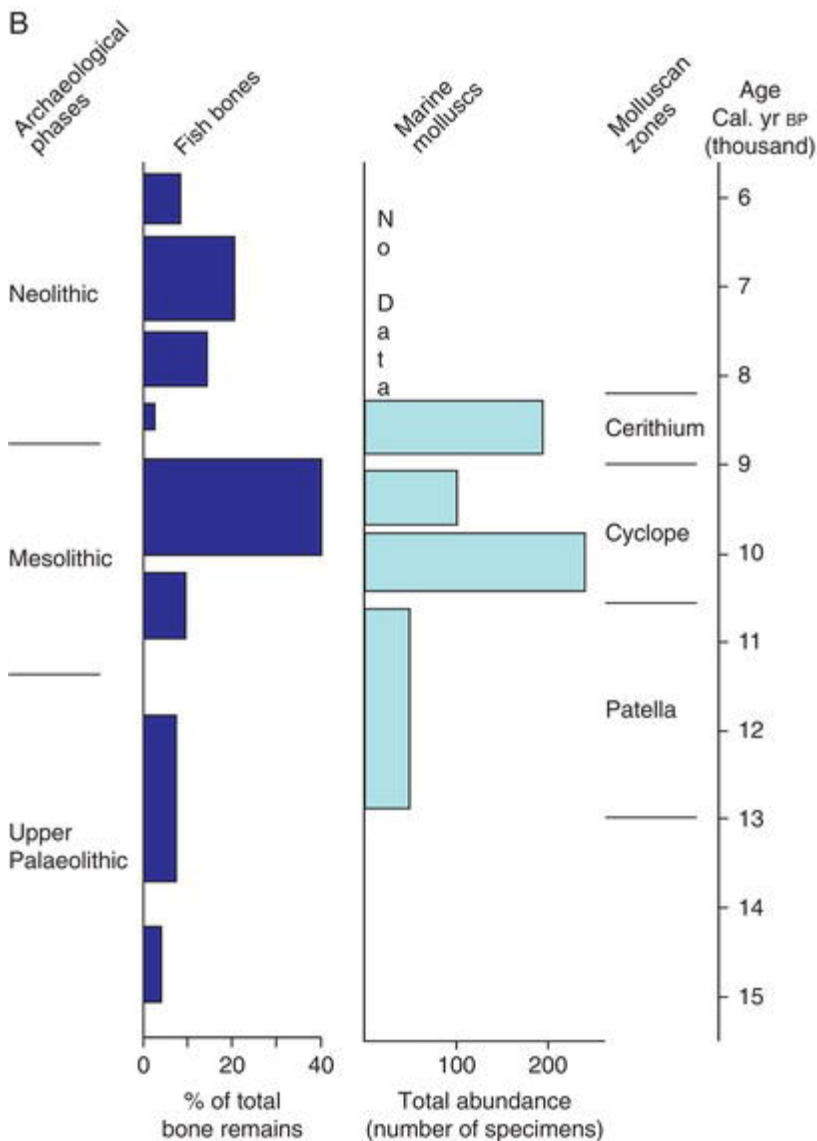
In other cases, the physical isolation brought about by rising sea levels led to cultural and economic retardation. Tasmania is a case in point. This continental island was joined to the Australian mainland during glacial low sea levels, but after 14 ka BP, the Bassian land bridge was severed (Jones, 1977). Although the indigenous Tasman aboriginal peoples possessed watercraft made of bark fibre and utilised coastal resources such as shellfish, they were unable to cross the stormy seas of the Bass Strait, today 250 km wide. The Tasman population did adapt to insular life and survived until their extinction following upon European contact, but ethnoarchaeological work has revealed that during the Holocene they 'lost' the use of many attributes including the boomerang, the concept of hafting and the habit of eating fish (Jones, 1977; p. 343). Nor did any of the technological innovations which occurred on the mainland after 14 ka BP reach Tasmania. It seems as if the rising post-glacial sea served to isolate Tasmanian society completely from mainstream aboriginal developments, producing an idiosyncratic culture analogous to the unique bird and animal evolutionary adaptations on many oceanic islands.

One of the best-documented adaptive responses to rising sea levels comes from Franchthi Cave in southern Greece. This site has an unusually long, unbroken archaeological sequence beginning before 15 ka BP and ending around 5.8 ka BP (Farrand, 1999). Today the cave entrance lies just a few metres above sea level on a rocky coast that can be reached only by boat. But as studies by Tjeerd van Andel showed, the situation was formerly different. At the start of the Holocene, the sea lay 2–3 km away, and most of the shore comprised a flat alluvial coastal plain. Seafood products, which had been largely absent from Pleistocene deposits in the cave, began to appear around this time and during the Mesolithic occupation assumed a major role in Franchthi's site economy (see Figure 4.2). Initially, the inhabitants were mainly content to collect shells, notably of *Cyclope neritea*, a small mollusc found on muddy nearshore and estuarine flats (Shackleton and van Andel, 1980; Shackleton, 1988). But they soon ventured further afield. Numerous bones of large fish such as tuna and the use of obsidian (black volcanic glass) from the Aegean island of Melos suggest deep-water fishing and sea-going boats. Around 9 ka BP, continuing sea-level rise appears to have drowned *Cyclope's* mud-flat habitat, since it is abruptly replaced in Franchthi's late Mesolithic levels by another mollusc – *Cerithium vulgatum* – characteristic of rocky shores.

Figure 4.2 Franchthi cave, Greece: A sea-level transgression (after van Andel et al., 1980; reproduced with permission of Maney Publishing); B sequence of marine resource usage (data from Payne, 1975; Shackleton and van Andel, 1980).

A





As well as reflecting changes in the proximity and hence the level of the sea, fish and shellfish remains from Franchthi may also offer a clue to the past productivity of Mediterranean waters. During the glacial-to-interglacial transition, the almost land-locked East Mediterranean basin received a large influx of freshwater. Initially, this came as meltwater from the retreating Eurasian ice sheets which cascaded from the Siberian lowlands via the Aral, Caspian and Black Seas into the Aegean (Grosswald, 1980). Later it originated from floods of a 'high and wild' stage of the river Nile around 10 ka BP (Adamson et al., 1980)

along with local increases in rainfall. This influx formed a freshwater 'lid' over the main body of saline seawater which became stagnant and **anaerobic**, allowing organic-rich sapropel muds to be deposited on the seabed (Williams et al., 1978; Fontugne et al., 1994; Rohling, 1994). The last East Mediterranean sapropel layer is dated to 9.5–6.5 ka BP, a period when levels of nutrients in seawater and marine productivity may have been higher than today. It was certainly the time when marine resources were most in evidence at Franchthi Cave. Further to the north, eustatic sea-level lowering had caused the Black Sea to be isolated from the world ocean, because the Bosphorus Straits which connect them are only about 60 m deep at present. In consequence, the Black Sea became a giant freshwater lake in glacial times, and the Bosphorus was transformed into an overflow channel down which poured the excess meltwater derived from ice sheets in its catchment. The present-day connection between the Black Sea and the Mediterranean was only established by rising world sea levels around 9.4 ka BP (Major et al., 2006).

The coastal zone was subject not only to long-term changes but also to short-term extreme events, such as storm surges. The tsunami which catastrophically inundated shorelines around the Indian Ocean on Boxing Day 2004 is a recent example which serves to remind us of the importance of rare, high magnitude events in Earth history. Major coastal flood events have occurred throughout the Holocene, leaving behind telltale sand layers containing marine microfossils. One such tsunami occurred in the North Sea around 8150 Cal. yr BP and left behind a flood deposit that can be traced along Scotland's northeast coast. This was triggered not by an earthquake but by a giant underwater mud-slide down the continental slope off Norway at Storrega, which created a wave more than 20 m in height (Bondevik et al., 2003). The ultimate cause of this turbidite-forming event was the accumulation of clastic debris derived from the Fennoscandian ice sheet which was then rendered unstable by rising post-glacial sea levels. Mesolithic communities living near to the shoreline of the North Sea must have thought that the world was coming to an end when this tsunami arrived. Certainly some human groups would have had little chance of survival, such as those living in 'Doggerland', a land now lost beneath the waves between England and the Netherlands. Events such as this may provide the basis for the ubiquitous flood myths found in many traditional cultures around the world. The most famous is the Noah's flood story of the Biblical Old Testament which originated in ancient Mesopotamia (Ryan and Pitman, 1998) and which may relate to folk memory of the post-glacial inundation of the coastal zone in southwest Asia.

Lake ontogeny and soil development

Pleistocene glaciation affected not only sea levels but was also a most effective lake-producing agency, for as the ice sheets retreated innumerable lakes were left behind in North America and northern Europe. Some, such as the Laurentide Great Lakes, were large and deep, but many more were simply small water-filled hollows, such as kettle-holes. Starting life as pools of infertile water underlain by till or rock, these young ecosystems have matured during the course of the Holocene. Lake maturation or **ontogeny** has classically been viewed in terms of a

series of stages, notably from an initial state of infertility or oligotrophy through eutrophy to extinction (see [Technical Box IV](#)). It is true that in small lakes, infilling by sediment progressively reduces water depth, while encroachment of plants from the shore reduces lake area. This succession from open water to reed swamp to fen is termed a [hydrosere](#) and like other ecological successions is often described in terms of a predictable orderly sequence of plant communities (see [Plate 4.2](#)). Donald Walker (1970) analysed the sediment and plant macrofossil records of cores taken at the edge of infilling lakes across the British Isles and found that they rarely show such an orderly sequence from open water to a dryland plant community. Placing 12 possible communities in sequential order, he found that most successional steps involved at least one seral stage being missed out. Not only were stages jumped, but in a significant number of cases, there was a reversion from a 'later' seral stage such as fen carr to an 'earlier' one such as reed swamp. This more complex vegetation sequence during lake infilling suggests that factors external to the lake, such as climatic variations, were at least as important in determining the ecological pathway as [autochthonous](#) ones were within the lake.

Other evidence which initially gave support to the stage model of lake evolution came from an increase in organic matter at the base of the sediment infill in many temperate lakes. Ed Deevey (1942) interpreted this increase as reflecting a sigmoid (or S-shaped) growth in biological productivity at the start of the Holocene. However, other palaeolimnologists such as Dan Livingstone (1957) questioned whether this organic carbon necessarily resulted from autochthonous production by algae. Organic matter can also originate from leaching or erosion of catchment soils and may, in any case, be diluted at times of high mineral inwash, such as during the terminal Pleistocene. John Mackereth working in the English Lake District suggested that lake muds could be better thought of as 'a series of samples of soils eroded from the drainage basin and deposited chronologically in the lake bed' (1966). If so, then it may be possible to reconstruct the postglacial development of temperate soils by analysing the chemistry of lake sediments. Mackereth hypothesised that the importance of the inorganic component of lake sediments would reflect the intensity of erosion in the catchment. In an actively eroding landscape, unweathered minerals such as Na (sodium) and K (potassium) would be brought into the lake. This occurred in Lake Windermere during the Late-Glacial stadial (13–11.7 ka BP) when pink minerogenic clays, low in organic matter, were deposited. On the other hand if the catchment was stabilised by soil and vegetation, runoff would instead carry dissolved ions and organic matter into the lake. Some of these would in due course be incorporated into lake bottom sediments, either directly or through aquatic organisms such as algae. The switch to organic sedimentation of this kind in Windermere began abruptly at the start of the Holocene (see [Plate 3.2](#)). We can therefore infer that in the protocratic phase of the Holocene, catchment soils were of base-rich, brown earth type, with accumulation of mull humus and nitrogen.

TECHNICAL BOX

Technical Box IV: Limnology

The study of lakes – or limnology – provides a good illustration of how physical, chemical and biological processes are interrelated within a single ecosystem (Hutchinson, 1957–75; O’Sullivan and Reynolds, 2003). Lakes may be large or small, shallow or deep. The water of lakes that are deep relative to their surface area will not be able to circulate completely for at least part of the year. In particular, only the surface water of temperate lakes is warmed up in summer, creating an epilimnion zone which is separated from the cold lower waters – or hypolimnion – by a sharp transition known as the thermocline (see [Figure 4.3](#)). In very deep, meromictic lakes such as Tanganyika, the thermocline persists all year round and the top and bottom waters never mix. More usually lake waters are separated for only part of the year, with seasonal overturn of top and bottom water. These are termed dimictic lakes. In the dark waters of the hypolimnion, photosynthesis may be negligible and the store of oxygen will become depleted, to create an anaerobic environment.

Lakes may be further characterized on the basis of their primary biological productivity. In those receiving ample nutrients from their catchments, algal productivity is relatively high and the lake [eutrophic](#). Nutrient-poor lakes, by contrast, are much less productive and are described as [oligotrophic](#). De-oxygenation of the hypolimnion is much more marked in eutrophic than in oligotrophic lakes, partly because of the continual ‘rain’ of dead organic matter derived from algal growth in the epilimnion falling towards the lake bed. The bottom sediments of eutrophic lakes therefore have a higher organic content than those of oligotrophic lakes. Under the reducing conditions typical of eutrophic bottom waters, phosphorus is released back into the lake water from the bottom

sediments, maintaining the nutrient supply.

Figure 4.3 Temperature and oxygen profiles of a stratified eutrophic lake in summer.

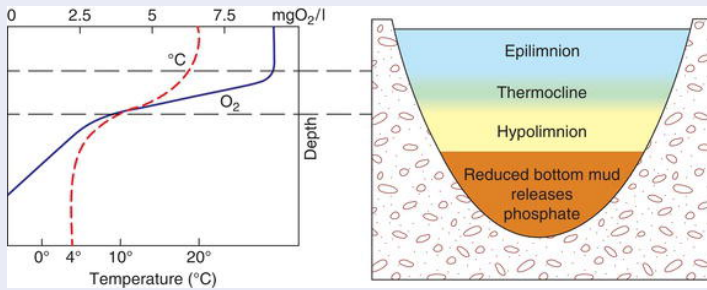


Plate 4.2 Blelham Tarn, English Lake District, with hydrosere sequence at lake edge.

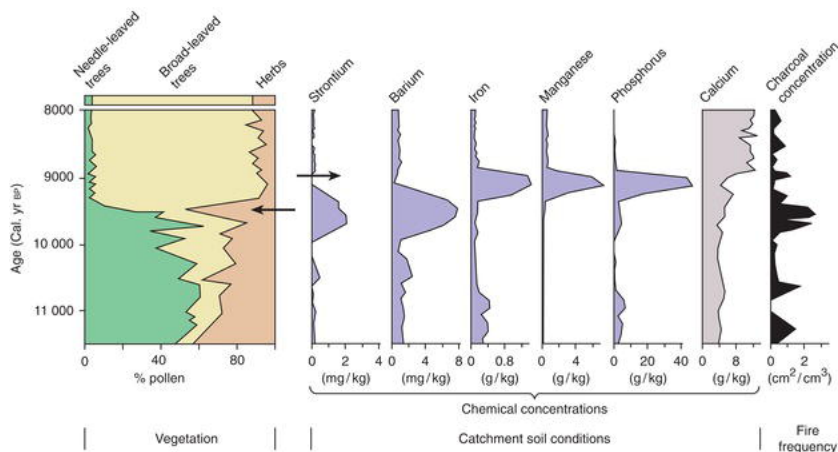


Although many lake ecosystems did experience high productivities during this protocratic phase, they were rarely maintained subsequently. Once limiting nutrients, such as P (phosphorus), had been released and used up, and once networks of ecological competition and predation were established, most lakes reverted to a mesotrophic or oligotrophic condition (Birks, 1986). The trend towards declining productivity and nutrient status continued during the mid-Holocene as soils became progressively leached. According to Svend Andersen's (1969) soil retrogression model, pedogenic changes from neutral to acid soils is an inevitable feature of these later stages of interglacials in many areas of moist, temperate climate (see [Figure 3.1](#)). In brown earth soils, litter from the mixed deciduous forest is rapidly broken down by soil organisms such as earthworms. However, with leaching of bases such as calcium, earthworms are replaced by

arthropods (insects and spiders), decomposition of litter is no longer complete, and acid mor humus begins to accumulate. Fe (iron) and Al (aluminium) are dissolved and moved down-profile to leave bleached sands over an impermeable iron-pan horizon – a typical podzolic profile. In some upland areas, this was the precursor to waterlogging and the formation of extensive blanket bogs (see Chapter 6).

Mackereth (1966) thought soil podzolisation should be recorded in lake sediments, specifically in the flux of iron and manganese. These elements are found in both mineral and organic soils, but they are relatively insoluble under oxidising conditions. On the other hand, under reducing conditions brought about by impeded drainage or the build-up of mor humus, Fe and Mn can be released from the soil in solution. Mackereth used the abundance of these elements in lake muds to reconstruct changing soil conditions. He argued, for instance, that with Mn being more soluble than Fe, the low Fe : Mn ratio in the sediments of Lake Windermere was a reflection of reducing conditions in the soils of its catchment, while the higher Fe : Mn ratio in Ennerdale Water was due to that lake's thinner, mineral soils. Other formerly glaciated upland areas experienced similar iron enrichment associated with podzolisation and leaching of soil humus, to create brown-water, **dystrophic** lakes. In Labrador, Canada, this occurred as boreal forest replaced shrub tundra between 8.4 and 7.8 ka BP (Engstrom and Wright, 1984). Palaeolimnology therefore indicates that in areas of acid bedrock, mineral soils were replaced by increasingly well-leached ones during the Holocene as Andersen's model predicted. The accompanying lowering of soil pH values had downstream consequences for streams and lakes, many of whose waters experienced a progressive, natural acidification over this time.

Figure 4.4 Early Holocene vegetation and soil changes reflected in a sediment core from Hungary (based on Willis et al., 1997). The arrows mark the main points of change, first in vegetation and subsequently in soils. Willis et al., 1997. Reproduced with permission of the Ecological Society of America.



In areas like the northeastern United States and central Europe which had supported boreal woodland in Late-Glacial times, soils evolved in the opposite

direction at the start of the Holocene – from podzol to brown earth. In eastern Hungary, for example, Kathy Willis used a combined analysis of pollen and sediment chemistry from the same lake core to examine how vegetation and catchment soils changed as the climate warmed (Willis et al., 1997). At the start of the Holocene, this area had supported an open coniferous forest-steppe, but at about 9.5 ka BP, the vegetation switched within no more than a century to mixed deciduous forest (Figure 4.4). Chemical analysis of the lake sediments shows that there was also release of chemical elements, such as Fe and Mn, which are typically associated with podzolic soils, followed by a sustained increase in calcium and reflecting the development of more alkaline, brown earth soils in the catchment. Significantly, these two sets of changes did not occur at exactly the same time, with the soil change lagging several centuries behind the arrival of deciduous trees. While there were undoubtedly a number of factors involved in the replacement of podzols by brown earth soils, such as wildfires (as reflected in a peak in charcoal), it is clear that here the post-glacial expansion of deciduous trees was a cause, and not a consequence, of changes in soil type.

The Holocene evolution of lowland temperate lakes was never so strongly limited by nutrient availability as were the acid, oligotrophic waters of the uplands. Lakes such as Lake Erie (United States/Canada) and Lough Neagh (Ireland) consequently became mesotrophic or eutrophic. It is apparent that the ontogeny of individual lakes depended on the particular combination of factors – geology, vegetation composition, soil development, lake morphometry, catchment:lake (z) ratio – that applied in each case. Some combinations produced pathways leading to dystrophy, others to eutrophy, yet others to infilling and extinction (Deevey, 1984).

The return of the forests

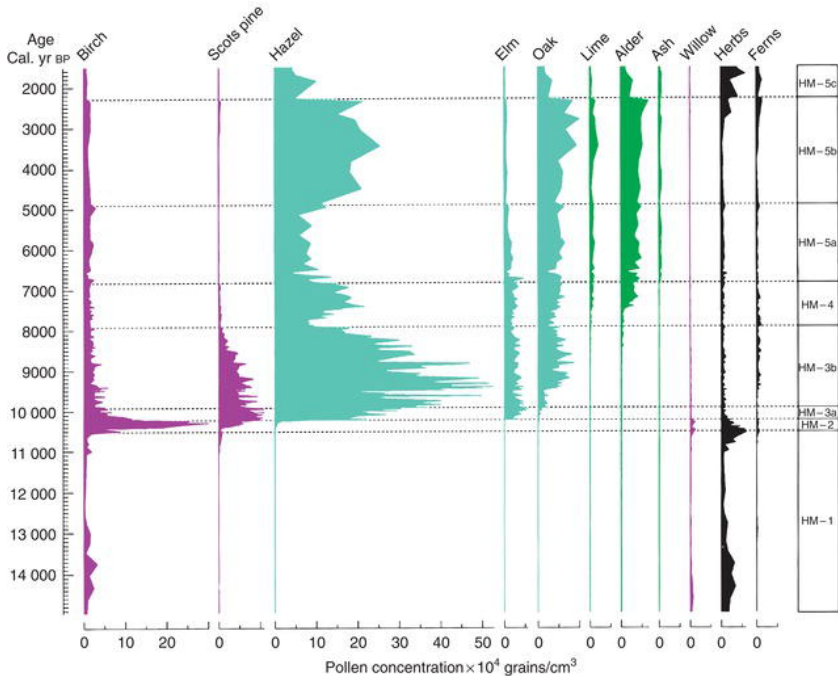
During the cold climate of the last glacial stage, mid-latitude deciduous and boreal woodland trees were located in lower latitude regions, often – as in the case of Europe – being forced into isolated refuges. Similarly, the tropical rainforests of Africa and Amazonia were reduced in extent and fragmented by climatic aridity. The subsequent return of forest ecosystems at the end of the Pleistocene and the beginning of the Holocene is one of the great stories of the Earth's recent natural history, and it is well documented by **palynology**. In particular, the many well-dated pollen diagrams from Europe and eastern North America make it possible to map the direction and rate of movement of both individual plant taxa and biomes across these two subcontinental land masses (Huntley and Birks, 1983; Webb et al., 1993; Prentice et al., 1996).

Europe

Perhaps surprisingly, Europe's glacial tundra-steppe and boreal forest remained

largely unchanged until almost 11.7 ka BP when viewed on a sub-continental scale. The restricted response of flora to Late-Glacial temperature fluctuations highlights the time lags that can occur between climate change and plant distribution. For example, despite experiencing temperatures close to those of today during the Late-Glacial interstadial, the British Isles remained a largely open landscape until the start of the Holocene. By this point in time, however, deciduous trees had begun to expand out of their southern refugia and ground conditions had been prepared by pioneering plant colonisers. Consequently, once climatic conditions permitted, there were few checks to a rapid spread of tree species across Europe, other than their own rates of dispersal (Hoek, 2001). Between 11.7 and 9 ka BP, tree species moved in to occupy suitable vacant land with much the same self-interested zeal as the miners of the '49 Gold Rush. The result was no less chaotic, with short-lived, localised associations of convenience formed between species – associations which often have no contemporary equivalent and hence no modern analogue. An example of the vegetation of this initial protocratic phase of the Holocene would be the sub-arctic birch–aspen woodland which existed adjacent to the retreating southern margin of the Scandinavian ice sheet. Studies of absolute pollen influx show a marked change during the protocratic phase, primarily because of a huge increase in vegetal biomass (see Figure 4.5). The concept of trees occupying phytologically empty land may therefore be a reasonable approximation to reality in early Holocene Europe.

Figure 4.5 ‘Absolute’ pollen diagram from Hockham Mere, Norfolk, England (after Bennett, 1983b). Bennett 1983. Reproduced with permission of John Wiley & Sons.



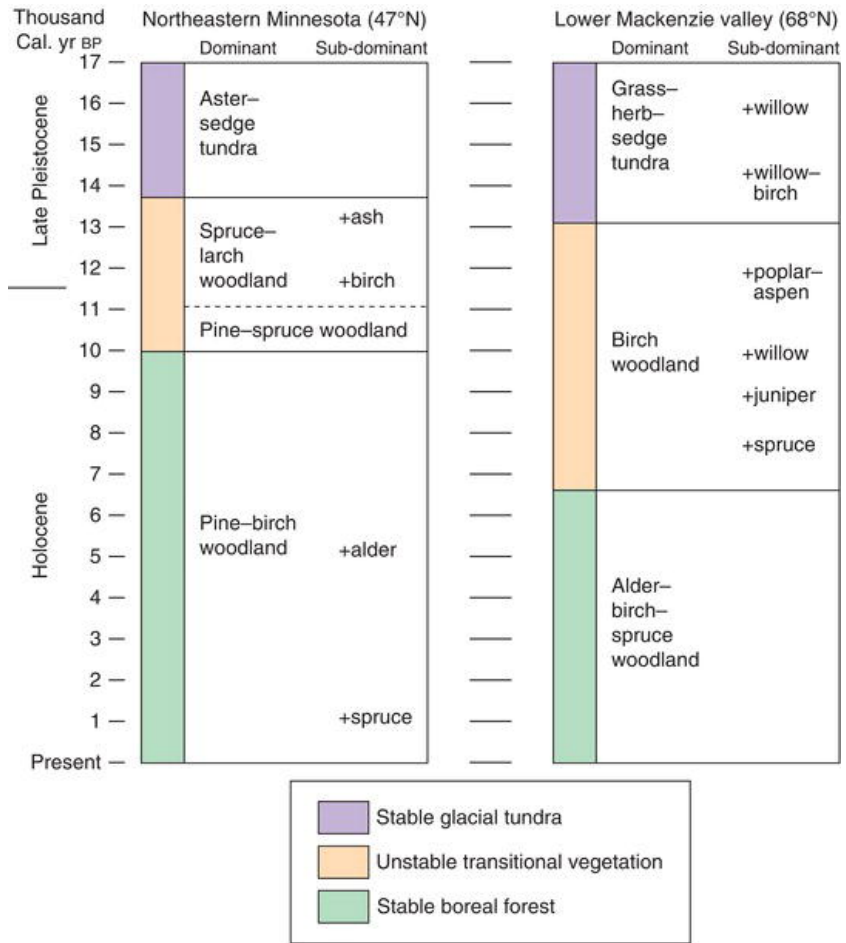
After two turbulent millennia of vegetation change, European phytogeography looked very different. The boreal forest had been pushed northwards to Scandinavia and northern Russia, tundra and steppe were all but removed from the scene, and the dominant vegetation type was now mixed deciduous forest. But if the distribution of plant formations had become essentially modern by 9 ka BP, their composition was not. This is immediately obvious from individual pollen diagrams, in which the characteristic feature is the continued arrival and rise to dominance of new woodland taxa (Giesecke et al., 2011). After the pioneer woods of birch and pine, the first deciduous trees to arrive in northwest Europe were hazel (*Corylus avellana*) and elm (*Ulmus*), both of which expanded without environmental constraints in less than 500 years (Bennett, 1983a). Later arrivals in the mesocratic forest were oak (*Quercus*), lime (*Tilia cordata*), alder (*Alnus glutinosa*) and ash (*Fraxinus excelsior*) (see Table 3.1). Other trees did not achieve their maximum extent until the declining or **oligocratic** stage of the Holocene, and some genera such as spruce (*Picea*) may still be expanding their range today (Huntley and Birks, 1983; p. 285).

Eastern North America

The return of temperate forest ecosystems began earlier in eastern North America, but in some areas finished later than it did in Europe. Two dominant palaeoclimatic and vegetational trends can be identified during the glacial-to-interglacial transition. The first was warming, which allowed temperate woodland species to spread northwards towards the retreating Laurentide ice sheet, pushing the boreal forest and tundra ahead of them. The second – desiccation – caused an eastward expansion of prairie grasses and forbs (leafy herbs). By the start of the Holocene, when Europe's vegetation still had a glacial character, the vegetation of the eastern United States already had a recognisably modern appearance (see Figure 4.6A). To the north, however, the wasting mass of the Laurentide ice sheet stubbornly refused to move out of the way. Ice, although much thinned, still covered most of what is now eastern Canada at the start of the Holocene, and it did not finally melt away until after 8.5 ka BP. As a result, northern vegetation formations were compressed against the ice sheet margins (Webb et al., 1993). Trees grew right up to the edge of the ice and led to the creation of buried forests, as at Two Creeks, Wisconsin, where the ice sheet temporarily readvanced over woodland around 13.8 ka BP. The creation of the boreal forest that now covers eastern Canada consequently only took place during early Holocene times. The time-transgressive nature of the transition from glacial to interglacial ecologies is reflected in the timing of vegetation changes from site to site (Ritchie, 1987). At the southern margin of the boreal forest zone, the unstable transitional period between tundra and forest lasted between c.14 ka and 10 ka BP, while at its northern limit, it started c.13.2 ka and was not finished until after 7 ka BP (see Figure 4.7).

Figure 4.6 Glacial-to-interglacial climatic changes in the North Atlantic sector (based on COHMAP Members, 1988); A environmental conditions as reconstructed from geological, pollen and ocean core records; B numerical simulation models of past atmospheric circulation patterns. Reproduced

America's boreal forest zone (based on data from Ritchie, 1985; Cushing, 1967).



Dry Mediterranean woodland

As with many arboreal vegetation formations, Mediterranean ecosystems experienced a contraction in their range during glacial periods of the Pleistocene. This was most acute in the Mediterranean basin where palynological research has yet to discover any clearly defined glacial refuge areas. Prior to c.14 ka BP, most of the northern littoral of the Mediterranean Sea appears to have been occupied by herb-dominated steppe in which *Artemisia* and chenopods were pre-eminent (Huntley and Birks, 1983; Rossignol-Strick, 1995). Evergreen oaks (*Quercus ilex*-type and *Quercus suber*) and olive (*Olea europaea*) were continuously present in small amounts throughout Iberia during the last glacial stage (López de Heredia et al., 2007; Carrión et al., 2010). The Atlantic cedar and *Pistacia* must also have had a refugium in the western Mediterranean. One of the few areas where evergreen

oak, *Pistacia* and olive are known to have survived together was the Southern Levant (Palestine, Jordan), although even here they formed a relatively minor part of the overall plant cover during the late Pleistocene (Van Zeist and Bottema, 1991).

The suggestion that some of the most important glacial refuges for Mediterranean flora lay in the eastern rather than in the western Mediterranean basin is given support by the earlier appearance of drought-adapted taxa pattern in the Holocene (see [Figure 6.9](#)). The expansion in tree species was by no means rapid, involving mean migration rates as low as 25 m/year in the case of manna ash (Huntley and Birks, 1983, p. 629). Furthermore, even after their appearance, typically Mediterranean taxa usually represented a minor component of the flora (Reille et al., 1996). The herb-steppe which surrounded most of the Mediterranean Sea during the late Pleistocene was replaced during the early and mid-Holocene by sub-humid forest, sometimes dominated by conifers, more usually by broad leaved deciduous trees (Van Zeist and Bottema, 1991; Lamb and van der Kaars, 1995; Reille et al., 1996; Sadori and Narcisi, 2001). The typically Mediterranean formations of xeric evergreen forests, shrub and heathland are only rarely represented in early and mid-Holocene pollen diagrams, even at low altitudes such as the littoral zone of southern Crete (Bottema, 1980). When present, they often indicate ephemeral plant associations, for instance, at Fos-sur-Mer in southern France, where *Phillyrea*, *Pistacia* and juniper occupied a geologically short-lived sand dune complex around 7 ka BP (Pons and Quezel, 1985).

The summer-dry woodlands of California and Australia appear to have been less severely restricted by glacial climatic conditions than those of the Mediterranean basin itself. Fossil forests of calcified tree stumps and other ¹⁴C-dated plant macrofossils indicate the existence of an Ice Age refugium in coastal California (Johnson, 1977), while the coastal zone of southern Australia has supported a diversity of Mediterranean-type habitats throughout the past 17 ka (Chappell and Grindrod, 1983).

Tropical forests

Vegetation in the tropics was as strongly affected by glacial climatic change as vegetation in the temperate zones. In montane environments such as the Andes, reduced temperatures led to the altitudinally zoned vegetation belts being depressed by over 1500 m (see [Figure 3.4](#)). At the start of the Holocene, as temperatures quickly rose, montane forest moved back up the mountains (Flenley, 1979b; Taylor, 1990). In lowland tropical environments, it was moisture availability as much as temperature which controlled late Quaternary vegetation history (Bonnefille et al., 1990; Markgraf, 1993; Marchant et al., 2004). During the period 20–15 ka BP, when active sand dunes extended into the present-day savanna zone, the lowland rainforests of Africa and South America had been reduced to isolated pockets by the arid climate (Elena et al., 2004).

At present, there are relatively few long pollen records with which to trace in detail the location of late Pleistocene lowland tropical forest refuges or the

pattern of postglacial recolonisation (e.g. Bush and Colinvaux, 1990; Wurster et al., 2010). At Lake Valencia in Venezuela, the major transition recorded during the last 15 000 years occurred between 12.5 and 9.7 ka BP (Bradbury et al., 1981). Before this, saline marshes occupied the lake basin and regional vegetation was dominated by open-habitat (i.e. non-tree) species. After it, Valencia was a deep freshwater lake, and the Holocene vegetation has remained one of savannas adjacent to the lake with evergreen and cloud forest on the surrounding mountains above 1000 m. Lowland forest is unlikely to have responded instantaneously to climatic change. Moist forest elements returned early in the Holocene at some tropical sites, often abruptly as at Barombi Mbo in Cameroon where the transition occurred around 10 ka BP (Elenga et al., 2004), but later at others, such as Lake Euramoo in the Queensland rainforest of Australia (c.7.4 ka BP). The African rainforest may have reached its maximum extent during the early Holocene, and Alan Hamilton (1982) suggested that a widely recorded rise in pollen of *Podocarpus* (southern yew) marks a change from moister to drier conditions in East Africa around 4 ka BP.

Factors affecting forest re-advance

During the glacial-to-interglacial transition, both individual taxa and whole plant formations shifted and usually expanded their ranges. But what were the factors that controlled the pattern of forest re-occupation of Europe, eastern North America and other regions at the onset of the Holocene?

Climate was obviously the ultimate determinant of ecological change during the Quaternary Ice Ages (Delcourt and Delcourt, 1987; Huntley and Webb, 1989). Less severe Glacial Age climatic conditions are probably the main reason for the greater diversity of tree taxa found in temperate North America than in Europe (Huntley, 1993a). However, it is by no means obvious that climate always directly controlled the rates of plant migration. In the North Atlantic region, ice cores and Coleoptera show an almost instantaneous rise in summer temperatures of around 7°C at the start of the Holocene (see [Figure 3.6](#)), a change confirmed by the early presence of thermophilous aquatic plants such as the bulrush (*Typha latifolia*). It seems unlikely that tree distributions were therefore in equilibrium with climate during the first 1500 years of the European Holocene (Birks, 1986), although climate seems to have acted as the main pacemaker for subsequent early and mid-Holocene changes in tree populations (Giesecke et al., 2011).

Carbon dioxide levels in the atmosphere affected vegetation not only via the greenhouse effect on climate but also directly through photosynthesis. Greater availability of CO₂ potentially means faster plant growth, while also favouring C₃ over C₄ plants (Bennett and Willis, 2000). As atmospheric CO₂ levels rose from around 200 ppm at the [LGM](#) to around 270 ppm at the start of the Holocene ([Figure 3.2](#)), this acted to ‘fertilize’ vegetation, increase overall NPP and biomass and change species composition towards C₃ plants, including C₃ grasses such as wild wheat and rice. Computer simulations suggest that the direct effect of changing CO₂ levels may have had as great an effect on forest structure as climate in eastern North America (Cowling, 1999). The effects may have been especially

keenly felt in the tropics, including in the vegetation zones on equatorial mountains in regions like East Africa (Street-Perrott et al., 1997).

Rates of dispersal and migration of individual tree taxa are obviously likely to have played a paramount role once the brake of climatic control was removed. In this regard, species whose seeds are wind dispersed, such as birch and elm, have an obvious advantage over those like oak which are reliant on other agencies, such as birds and squirrels, for carrying their seeds away from the parent tree. The growth rate and age of first seed setting for individual trees are other important determinants of the speed of migration (see Table 3.1). Mighty oaks from tiny acorns may grow, but they do not do it overnight. The rates of migration of individual taxa at the start of the Holocene can be calculated from ^{14}C -dated pollen diagrams across Europe and eastern North America. These reveal that in Europe, several opportunistic genera, notably birch, pine, alder and hazel, expanded at rates of 1–2 km/year over periods of 500–2000 years. That is, their seeds would have required dispersal over 10–20 km even if the trees were able to mature and fruit in only 10 years (Huntley and Birks, 1983; p. 632). In North America, trees have maximum migration rates of around 0.5 km/year (Davis, 1976), suggesting that other factors such as climate and competition with the existing spruce forest operated to slow down the return of deciduous trees. The American Jack pine (*Pinus banksiana*) provides an example. Today, it is found in the northern Great Lakes region and in central Canada, but pollen maps show a glacial refuge in the southeast United States, from which it spread northwestwards at an average rate of 350–500 m/year between 15 ka and 9 ka BP (Davis, 1976). Outlying populations of Jack pine still exist in New England as relics of its migration through the area 12 000 years ago.

Obviously, the time of arrival of individual taxa depended not only on the rate of migration but also on the position from which they started their re-advance. In Europe and the Mediterranean basin, principal **glacial refuge areas** have been identified in Iberia, Italy, southeast Europe and Turkey on pollen evidence (Huntley and Birks, 1983, p. 627; Bennett et al., 1991). It was, for example, easier for pine to reoccupy northern Europe than it was for beech (*Fagus*), not only because of its faster rate of migration but also because its spread started from western Ireland and Poland, rather than from Italy and the Balkans (Bennett, 2008). By combining pollen and plant macrofossil analysis with plant genetics, it has been possible to show that trees such as beech and oak survived the last glacial period in multiple refuge areas and how some populations have greatly expanded their range during the postglacial period, while others experienced only a limited expansion (Magri et al., 2006; López de Heredia et al., 2007). Not all glacial vegetation refuges were as far removed from their modern distribution as they were in Europe or North America. In New Zealand, for example, southern beech (*Nothofagus*) was restricted to locally favourable sites in an otherwise open glacial landscape. With the rapid terminal Pleistocene warming which characterised most of the southern hemisphere, the forest re-established itself and within 300 years had replaced scrub and grassland (McGlone et al., 1993). This illustrates how rapidly plants can respond to climatic changes in the absence of soil or distance constraints.

Did lack of **soil development** restrict the re-advance of arboreal elements, especially of deep-rooting, nutrient-demanding species? In those areas formerly

under ice, deglaciation left raw, undeveloped soils which favoured open vegetation (Pennington, 1986). In New England, for instance, sedge-dominated tundra is recorded at Rogers Lake between the retreat of the ice and c.13.5 ka BP, while at the same time spruce forest lay close by in the zone just outside the former glacial limits (Davis, 1976; Watts, 1983). But although soil factors determined the speed and character of the early stages of plant succession around some ice sheet margins, in other cases, the arrival of new vegetation affected soil type more than the other way round (Willis et al., 1997). Soils may, in any case, have been less important in this protocratic phase than they were to be in the succeeding mesocratic and oligocratic periods. Regional soil differences, for example, were to influence strongly the polyclimax nature of Europe's mixed deciduous forest zone during the mid-Holocene.

Competition between species certainly influenced early Holocene ecological succession. In northern Europe, the rapidly dispersed silver birch (*Betula pendula*) prepared the ground for later arrivals with which it was then unable to compete. In particular, it was shaded out by the dense forest canopy created by trees such as elm. The slower expansion of taxa such as oak and the failure of beech to establish itself as a major forest tree in parts of northern Europe until the later Holocene may also have been due to competitive exclusion.

Physiographic barriers such as the Alps and the North Sea hindered the free movement of forest taxa. On the other hand, the generally south to north drainage orientation in Europe led to long-distance water dispersal of hazelnuts and acorns that would otherwise have spread more slowly. In eastern North America, there are no major mountain barriers or epicontinental seas, although a major obstacle, no longer extant, did exist in the form of the Laurentide ice sheet. The effect of this on forest history can be seen clearly in the case of the spruce during the terminal Pleistocene. Rising temperatures allowed spruce forest to occupy much of the former periglacial zone of northeastern and midwestern United States. By the start of the Holocene, however, this boreal forest tree was being overtaken from behind by later invaders such as pine and oak, but was itself unable to move northwards on account of the residual ice mass occupying eastern Canada (see [Figure 4.6A](#)). Squashed up against this physiographic barrier, spruce experienced a catastrophic population decline which is dated in pollen diagrams to around 11.7 ka BP and which, in some cases, occurred over as short a time as 50 years (Watts, 1983; p. 308). The ice sheet which prevented forest invasion in North America ironically had the opposite effect in other parts of the world. The continued existence of the Laurentide ice sheet kept eustatic sea levels low and meant that, for example, the British Isles were not separated from the European mainland until after 8.5 ka BP, that is, after the main wave of tree migrations was complete. In consequence, the native British flora lacks only oligocratic trees such as spruce (*Picea abies*). On the other hand, once isolated, tree populations on islands are vulnerable to local extinction, such as that which occurred in Ireland with pine during the late Holocene (Bradshaw and Browne, 1987).

Other factors, too, played their part. They include patterns of animal browsing, woodland management by hunter-fisher-gatherer groups, disease and fire. Frans Vera (2000) hypothesized that, before the advent of agriculture, browsing by large herbivores like aurochs (wild cattle), European bison, deer and

tarpan (wild horse) acted to maintain a mosaic landscape of grassland, parkland and small wooded copses and hence prevented the dominance of closed-canopy deciduous forest. Fraser Mitchell (2005) tested this idea by exploiting the fact that there were few, if any, large herbivores in Ireland during the early Holocene (Table 4.1). By showing that there were no appreciable differences between Irish and mainland European pollen records at this time, he was effectively able to falsify the Vera hypothesis. However, it is important to note that different factors operated over different spatial scales. Alder (*A. glutinosa*), for example, became established as a major European forest tree because of the existence of waterlogged soils at the base of local slope catenas, where it was able to compete most successfully.

Table 4.1 Terrestrial mammals of Britain and Ireland during the Late Glacial and Holocene

Species	Late Glacial	Early Holocene		Present day	
		Britain	Ireland	Britain	Ireland
Bison	x				
Cave bear	x				
Giant deer ^a	x				
Mammoth	x				
Woolly rhinoceros	x				
Lion	x				
Lemming	x				
Arctic fox	x				
Tundra vole	x				
Elk	x	x			
Horse	x	x			
Reindeer	x	x			
Wolf	x	x	x		
Brown bear	x	x	x		
Lynx	x	x	x		
Otter	x	x	x	x	x
Stoat	x	x	x	x	x
Blue hare	x	x	x	x	x
Pygmy shrew	x	x	x	x	x
Fox	x	x	x	x	x
Badger		x	x	x	x
Pine marten		x	x	x	x
Red deer		x	x	x	(x)
Wood mouse		x	x	x	x
Wild cat		x	x	x	
Polecat		x	x	x	
Wild boar		x	x	(x)	
Common shrew		x		x	x
Mole		x		x	x
Hedgehog		x		x	x
Red squirrel		x		x	x
Brown hare		x		x	x
Roe deer		x		x	
Weasel		x		x	
Field and water vole		x		x	
Yellow-necked mouse		x		x	
Water shrew		x		x	
Aurochs		x			
Beaver		x		(x)	
Rabbit				x	x

^aIrish elk or *Megaloceros*

(x) indicates reintroduction

Source: Adapted from Grigson in Simmons and Tooley (1981) and other sources

An almost universal characteristic of this period of dynamic vegetation change is that species behaviour was strongly individualistic (Delcourt and

Delcourt, 1987). Particular genera or species may be co-associated as part of a vegetation formation at the present day, but at times of climatic transition, each one moved on its own, and some faster than others. Forest communities did not move *en bloc* at the start of the Holocene but were created and destroyed as individual taxa arrived and departed. This applies as much to tropical vegetation formations such as the montane rainforest of Papua New Guinea (Flenley, 1979a) as it does to the mixed deciduous forest of the temperate zone (Davis, 1976). The variable migration rates and directions for different taxa created ephemeral plant communities which often lack a modern analogue – such as the combination of spruce, oak and ash which existed in the midwestern United States at the end of the Pleistocene and which do not now occur sympatrically (Watts, 1983).

The ecology of Mesolithic Europe

During the late Pleistocene, Europe's mid-latitude tundra had supported large herds of reindeer (*Rangifer*), horse (*Equus*) and other herbivores (see Chapter 3). This fauna was displaced northwards at the start of the Holocene as temperate forests returned, and in its place came new woodland animals such as red deer (*Cervus elaphus*), aurochs (*Bos primigenius*) and wild boar (*Sus scrofa*). As resources for human exploitation, these animals were more dispersed and less visible in the forests than had been the concentrated and easily culled fauna of the Late-Glacial tundra. But if animal resources became less accessible after c.11.7 ka BP, plant resources changed in exactly the opposite way. Mixed deciduous woodland is a biome of high, if strongly seasonal, primary productivity and contains several hundred potentially edible plant species. These range from tree products such as hazelnuts through berries and fruit to fungi and bracken rhizomes. As Ray Mears and Gordon Hillman (2007) have reminded us, these could be collected for much less expenditure of effort than their hunted meat equivalent. Any child more than four years old could gather sufficient to feed itself, and in habitats like the stone pinewoods of southern Europe, they would collect enough for the rest of the family too (see [Plate 4.3](#)). The consumption of plant foods generally leaves fewer traces than the bones left after animal hunting, and it is consequently difficult to make quantitative estimates of the relative dietary importance of each at individual archaeological sites. Nonetheless, the fact that bone remains are well represented at [Mesolithic](#) sites indicates that Europe's early and mid-Holocene population was omnivorous rather than vegetarian.

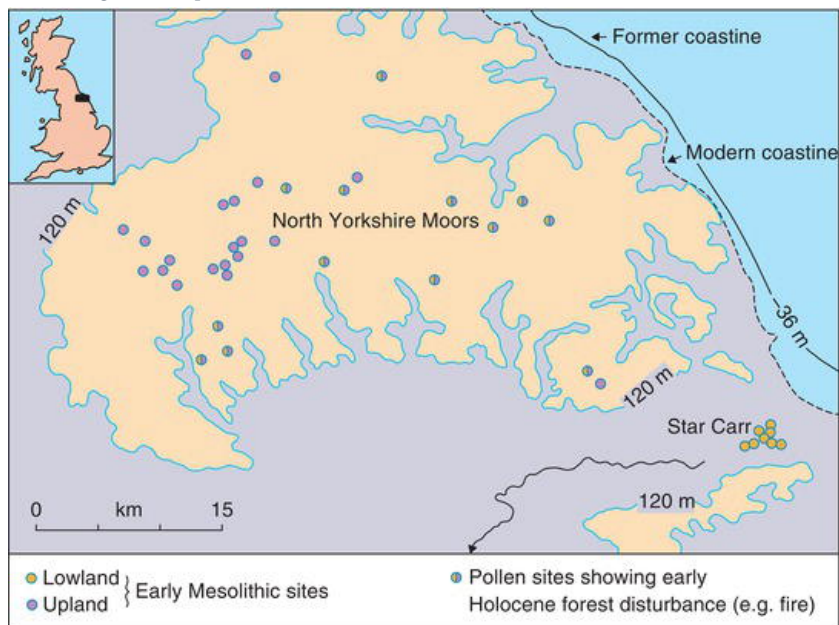
Plate 4.3 The Mediterranean stone pine (*Pinus pinea*); its kernels have a protein value two-thirds that of lean steak.



The seasonal rhythm of plant growth and animal movement in temperate woodland ecosystems was strongly to influence the food schedules and lifestyles of Mesolithic h-f-g groups. In northern England, for example, Mesolithic sites are found in two clear altitudinal bands: one within 100 m of modern sea level, the other on higher ground between 250 and 500 m (Jacobi et al., 1976). Among the lowland sites is Star Carr, the scene of famous archaeological investigations by Grahame Clark between 1949 and 1953 (Clark, 1972). His were among the first excavations to be as concerned with the site economy and environment, as revealed by bones, seeds and pollen grains, as with the typology of stone tools. Study of elk (*Alces*) and red deer antlers found at Star Carr initially suggested that the site was most intensively occupied in winter and early spring. Clark proposed that in summer most of the inhabitants moved away with the red deer to the higher ground of the North Yorkshire Moors, which were then covered by open birchwood and hazel scrub. This seasonal transhumant cycle would, he suggested, have allowed deer and their hunters to avoid the winter snows on the high ground and the insect-ridden lowlands in summer. Subsequent researchers (e.g. Caulfield, 1978; Andresen et al., 1981) have cast doubt on Clark's original model of Star Carr as a winter base camp; the antlers, it is thought, were brought onto site by hunters rather than being shed by the deer locally. Re-analysis of the animal bones by Tony Legge and Peter Rowley-Conwy (1988) instead led to the interpretation of Star Carr itself as a late spring-summer hunting camp. Nonetheless, it is likely that Mesolithic bands followed a seasonal cycle involving 'fission and fusion' similar to that observed in modern hunters and gatherers. Bands may have come together at lowland base camps to pass the winter – a season of climatic stress and food shortage – but split up into smaller family-sized units across their summer territory. This model is given archaeological support from site types, lowland Mesolithic sites being generally larger and more substantial (c.20 people) than their upland equivalents (c.5 people) (see [Figure](#)

4.8).

Figure 4.8 Pattern of Mesolithic settlement and early Holocene vegetation disturbance in North Yorkshire, England (adapted from Jacobi, 1978; Simmons and Innes, 1985).



As the relatively open birch, hazel and pine woods of the protocratic phase in Europe gave way to the closed-canopy mesocratic forest, open habitats became increasingly scarce. Openings in lowland forests were found mainly near to water, either on the coast or next to lakes, marshes and rivers. Star Carr itself comprised a birchwood platform at the swampy edge of a lake, now infilled (Cloutman, 1988, Conneller et al., 2009), the waterlogged conditions being responsible for the excellent preservation of organic remains at the site. Wetlands are resource-rich environments, whether for edible water plants such as cress and water lily (Clarke, 1976) or for fowling and fishing. The latter was especially important in Ireland, which had been cut off by rising sea levels before many woodland animals had time to arrive (see Table 4.1). Compensation for Ireland's depauperate mammal fauna came from its many lakes and rivers. For example, along the Lower Bann river between Lough Neagh and the sea, the bones of large numbers of migratory fish, notably salmon and eel, have been recovered from Mesolithic excavations (Woodman, 1978, 1985).

In upland environments, openings in the forest lay mainly towards the upper altitudinal limit of tree growth. Here fire provided an alternative method of deforestation to ring-barking or felling with Mesolithic tranchet stone axes. Selective burning is a traditional technique of environmental management practised by hunters and pastoralists throughout the world (Mellars, 1975). The new vegetation growth after a fire increases grazing and browsing potential, and the number of deer or cattle that can be supported responds accordingly. Fire may

also have been used to encourage production of acorns – a potentially nutritious food source (Mason, 2000). Charcoal in soil and peat profiles and palynological evidence for a reduced forest cover in areas such as the southern Pennines and Dartmoor suggest that recurrent burning of upland vegetation took place during the Mesolithic (Simmons and Innes, 1985; Tallis and Switsur, 1990; Caseldine and Hutton, 1993). It is probable that much of this resulted from deliberate firing, as this ecotone experienced substantial and long-continued Mesolithic occupation.

The early Holocene in the tropics

Early research on the Quaternary in the tropics recognized that there had been wetter and drier climatic phases in the past. However, in the absence of independent dating, these phases were fitted around then-prevailing glacial–interglacial frameworks that had been developed for higher latitudes. Around 1960, matters began to change. One reason for that change was the discovery at Olduvai Gorge in Tanzania of hominin bones dated radiometrically to over 1.5 million years old, indicating that the human race had its origins on the African continent. A second change, more directly relevant to this chapter, came with the application of ^{14}C dating to fossil lake beds in the Sahel and the East African rift (see [Plate 4.4](#)). These former lakes, interpreted as the product of a wetter ‘**pluvial**’ climate, had previously been correlated with the glacial phases of Europe and North America and might have been expected to produce calibrated ^{14}C ages around 20 000 years old. However, when the French geologist Hugues Faure sent off samples of freshwater carbonate and diatomite from lake basins in the Sahel, he received back much younger ^{14}C ages, between 10 300 and 7700 years old after calibration (Faure et al., 1963). Not surprisingly, this encouraged further research work which similarly yielded ^{14}C dates in the first half of the Holocene. Karl Butzer and others (1972) were able to put together the results from a number of East African lakes and show that their water levels had fluctuated in concert over the past 15 000 years. Non-outlet lakes were clearly an important source of palaeoclimatic information in low-latitude regions. By the late 1970s, so many lake basins were under study and so many ^{14}C dates produced that it was necessary to centralise them in a global databank, compiled by Alayne Street-Perrott and held initially at Oxford. In this database, lakes are classified as at high, intermediate or low levels at 1000-year intervals within the time period for which reliable ^{14}C chronologies exist (Street and Grove, 1979; Street-Perrott and Roberts, 1983). From before 10 ka to c.5.5 ka BP, there was a truly massive extension in the zone of high-level lakes and water surplus, not only across the semi-arid Sahel but into the Sahara desert itself (Hoelzmann et al., 2004), labelled the African Humid Period.

Plate 4.4 Early Holocene lake sediments (10 m thick) from Taoudenni in the central Sahara, subsequently eroded by the wind to form pillars. Photo credit: Nicole Petit-Maire.



Saharan palaeoecology

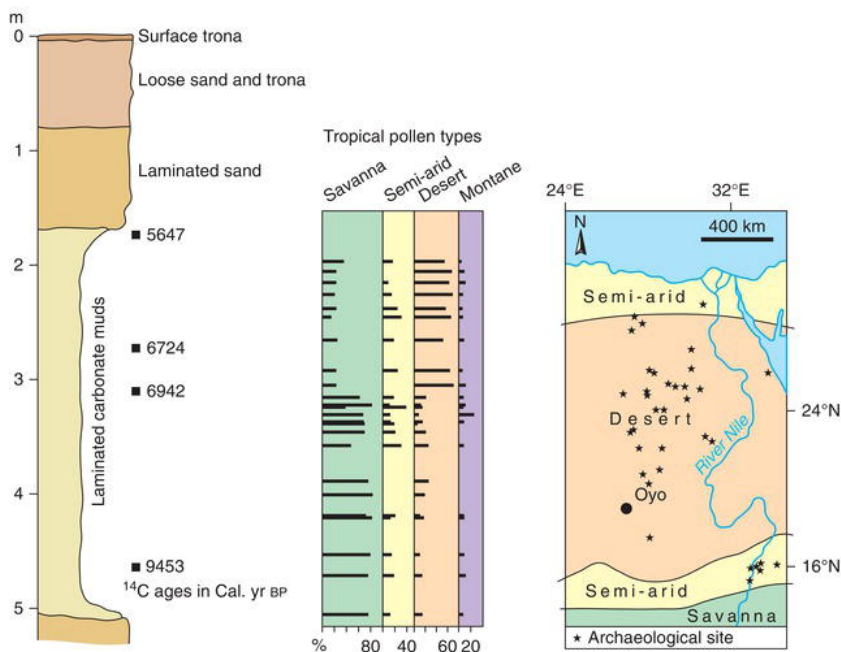
Until almost the time of Egypt's pyramid builders (4.5 ka BP), the ecology of the Sahara was not the present hyper-arid wasteland but a veritable land of lakes. The presence of permanent freshwater is indicated not only from mollusc shells, diatoms and laminated lake sediments but also – and more spectacularly – from bones of large aquatic fauna. From northern Mali, where rainfall now averages only 5 mm/year, have come finds of crocodile (*Crocodylus niloticus*) up to 3 m long and hippo (*Hippopotamus* sp.) (Petit-Maire and Riser, 1983). These and other fauna typical of savanna ecosystems required not only water but also plant life in order to exist.

Pollen grains are generally poorly preserved in arid environments, but several sites covering the full width of the Sahara have now been successfully investigated. In the west (Senegal, Mauritania), relict forests of the humid Guinean type confirm pollen evidence of a southerly vegetation shift in vegetation zones since mid-Holocene times (Lézine, 1989). A strong mid-Holocene augmentation in pollen of sub-humid Sudanian vegetation is also recorded in and around the Lake Chad basin, now in the semi-arid Sahelian climatic and vegetation zone (Hoelzmann et al., 2004; Lézine et al., 2011). Some of the most startling evidence for the African Humid Period comes from Oyo in the heart of the hyper-arid eastern Sahara (Ritchie and Haynes, 1987). In pollen spectra dating from before 9500 to c.7000 Cal. yr BP at this former lake, most non-local pollen was of Sudanian type, including grains of tropical taxa such as *Hibiscus* which are produced by plants in only small quantities and are poorly dispersed (see [Figure 4.9](#)). During this period, Oyo – then a deep, stratified lake – must have been surrounded by savanna vegetation similar to that now found 500 km to the south. Between 7000 and 5700 Cal. yr BP, the lake became shallower and acacia-

thorn and then scrub grassland replaced the sub-humid savanna. Finally, around 5100 Cal. yr BP, the lake dried out, aeolian activity recommenced, and vegetation disappeared except in wadis and oases.

The Saharan desert which had been considerably more extensive than at present during the late Pleistocene (20–15 ka BP) effectively did not exist during most of the early Holocene. The savanna which replaced most of the present arid zone was relatively hospitable to human life, and there is abundant archaeological evidence including some superb rock paintings (see [Plate 4.5](#)) that the human ecology was then radically different from that which could exist in the Sahara today (Kuper and Kröpelin, 2006) ([Figure 4.9](#)). Of course, the environmental resources varied considerably across this huge area. In the Egyptian Sahara, for example, early Holocene game was dominated by gazelle (*Gazella dorcas*) and hare (*Lepus capensis*) and lacked the diversity of the savanna fauna then found in northern Mali (Gautier, 1980). For the most part, human subsistence patterns varied according to the resources available, but the variety of regional specialities would have impressed even the best French chef. Freshwater turtles, molluscs and fish were an important component of diet in the southern Saharan Neolithic, which has been appropriately labelled the Aqualithic culture. On the northwestern edge of the Sahara, where conditions were semi-arid rather than humid, the distinguishing feature of the cuisine was its fondness for escargots – edible land snails. The so-called Escargotières of the Capsian culture are small shell middens that were occupied occasionally but repeatedly for up to two or three millennia during the early Holocene (Lubell et al., 1976). Each of the snail species exploited occupies a separate ecological niche that has a somewhat different seasonal ethology so that foraging would have rotated from site to site, allowing the snail population to recover in between times, preventing over-predation. Although whole shells averaged 25 000/m³ at excavated sites, the image of a population feasted on this culinary delicacy is probably a false one. In truth, the snails were probably placed in the pot along with meat and wild plants, and on their own accounted for only a limited proportion of the total dietary base.

Figure 4.9 Pollen diagram from Oyo (adapted from Ritchie et al., 1985) along with modern vegetation zones and the distribution of archaeological sites during the peak of the ‘green Sahara’, 9.0–7.3 ka BP (data from Kuper and Kröpelin, 2006).



Both of these economies fall within the province of hunting-fishing-gathering, but a third adaptation to Holocene Saharan environments involved a more substantial manipulation of food resources. In what is now the Egyptian desert, Fred Wendorf and his colleagues (Wendorf and Schild, 2001) discovered and excavated Neolithic sites which included some domestic species, notably cattle. The beef eaters of this area lay adjacent to the major southwest Asian nuclear hearth of plant and animal domestication, and they represent the likely origin of the cattle-based cultures which were to spread throughout Africa during the course of the Holocene.

Plate 4.5 Mid-Holocene rock painting of a giraffe head and hunters from Tassili in the arid heart of the Sahara desert. Under the former wetter climate, savanna wildlife was common, and so too were people hunting them. Photo: James Wellard. Reproduced with permission of Sonia Halliday Photographs.



Early Holocene climates: Pattern and process

In contrast to the tropics, in regions such as temperate Europe it is not easy from pollen or other proxy data to identify unambiguously major climatic shifts since the last deglaciation. Certainly, cooler/wetter or warmer/drier phases of the Holocene were recognized by Blytt and Sernander (see [Technical Box V](#)). Similarly, a thermal optimum (9000–5500 Cal. yr BP) was identified by Iversen and others from the former extension of plants such as the water chestnut (*Trapa natans*) and animals such as the pond tortoise (*Emys orbicularis*) north of their present climatic limits (Sommer et al., 2007). But these changes were minor in comparison with the fluctuations in climate that affected Europe during the millennia prior to 11.7 ka BP. It was therefore presumed that ‘from at least 7000 BC, and possibly earlier, the worldwide climate had been essentially the same as that of today’ (Raikes, 1967).

TECHNICAL BOX

Technical Box V: Subdividing the Holocene

How should we divide up Holocene time? In Europe, main basis for subdivision has traditionally been

based on changes in vegetation and inferred climate. One especially influential framework has been the Blytt-Sernander model, originally based upon peat stratigraphy (Charman, 2002). Blytt and Sernander used phases of faster and slower peat growth to divide the Holocene into phases of inferred wetter and drier climate (see Table 4.2). With the subsequent development of **palynology**, pollen zones were fitted in with the same scheme. While recent work has supported the idea that phases of peat growth were caused by climatic changes, deficiencies have been found in their original scheme. Most ombrotrophic bogs in fact show a more complex sequence of recurrent surfaces from humified to unhumified peat than were recorded by Blytt and Sernander (see Figure 6.2). Furthermore, with the application of ^{14}C dating it has become clear that Holocene vegetational changes were time-transgressive. The Blytt-Sernander scheme is consequently falling into disuse, although in a colloquial sense terms such as the ‘Atlantic oakwoods’ are still used.

Table 4.2 Blytt–Sernander stages for the European Holocene

Period	Peat type	Inferred climate	Pollen zone	Approximate age (ka BP)
Sub-Atlantic	Unhumified peat	Cool/wet	VIII	Present 2.6
Sub-Boreal	Humified peat + pine stumps	Warm/dry	VIIb	2.6–5.7
Atlantic	Unhumified peat	Warm/wet	VIIa	5.7–7.8
Boreal	Humified peat + pine stumps	Warm/dry	V/VI	7.8–10.5
Pre-Boreal	Hydrosere peat	Cool/dry	IV	10.5–11.7

Instead it has become common currency to split the Holocene epoch into early, middle and late phases. It has now been proposed to formalise these periods in line with the same conventions that apply to the whole of the geological timescale. Walker et al. (2012) recommend using an abrupt climatic event at 8.2 ka BP to divide early from middle Holocene, and a

similar – although less well-defined – event at 4.2 ka BP to separate the middle from late Holocene. Any attempt at formal subdivision encounters the same arguments already outlined for defining the start of the Holocene ([Technical Box I](#)), namely should boundaries be synchronous or time-transgressive, and is it possible to find global strato-type sections? The 8.2 ka event has the advantage of being widely recognized across the northern hemisphere, including – importantly – the Greenland ice cores. On the other hand, climatic conditions were not any different before and after this short-lived cold, dry excursion, so diminishing its value in terms of longer-term global change. The inferred 4.2 ka BP climatic event is even more problematic in terms of global climato-stratigraphy. It is not present in the Greenland ice core record, its causes are not known and it is not even clear if it was synchronous in different parts of the planet. These difficulties highlight the problems of applying geological rules developed for ‘deep time’ to more recent time periods. It is something which becomes even more obvious as we approach the present-day (see [Technical Box IX](#)).

It is now known that climatic stability did not characterize the Holocene everywhere in the world. As noted earlier, in the tropics and sub-tropics, the climate 9000 or 6000 years ago was substantially different from that of the present day (COHMAP Members, 1988). These differences are clearly manifested in palaeoecological and geological records and can only realistically be explained by shifts in atmospheric circulation patterns. This should cause us to re-examine more closely the climatic record of temperate regions. Was the climate of Holocene Europe really as stable as has been often envisaged? Or was the dominant mid-latitude vegetation formation of mixed deciduous forest simply ecologically insensitive to climate?

Climatic interpretation of proxy data is rarely straightforward. Former high lake levels, for example, could be the product of reduced evaporation at times of low temperatures rather than any absolute increase in precipitation. Fortunately, there is no such problem of interpretation with the enlarged early Holocene lakes in Africa. They formed when temperatures were similar to today's, so high water levels can only have been caused by higher rainfall. Calculations using both water

and energy balance approaches indicate that rainfall in East Africa and the Sahara increased by between 150 and 400 mm per annum (Street-Perrott et al., 1991). What was the source of this enhanced moisture? The palaeoenvironmental record is clear in indicating that the zone of tropical convectional rains moved northwards and broadened during the early Holocene. Mid-latitude depressions may have interacted with tropical air masses over some parts of the western Sahara and over high ground such as the Tibesti Plateau, but were of only secondary importance (Nicholson and Flohn, 1980). A northerly shift in the Inter Tropical Convergence Zone (ITCZ) and its associated rains is indicated by a poleward shift in the Sudanian–Sahelian vegetation belts and by early Holocene lakes that were larger, deeper and more permanent towards the southern side of the Sahara.

An equally important indicator comes from changes recorded outside Africa. If the ITCZ moved north over the African continent, it would also have been expected to do so over the Indian Ocean sector which today forms a linked part of the same monsoonal circulation system. Comparable evidence for enhanced summer rainfall does indeed occur in the Arabian and Rajasthan (northwest Indian) deserts, in the form of fossil lake beds, isotopes from cave speleothems, palaeosols and pollen indicative of savanna vegetation (McClure, 1976; Burns et al., 2001; Fleitmann et al., 2003; Parker et al., 2006). Indian Ocean deep-sea sediment cores also record stronger upwelling of the Arabian coast and pollen of tropical origin (Sirocko et al., 1991; Overpeck et al., 1996). There is even evidence of high lake levels and changed vegetation composition as far north as the plateau of Tibet (Gasse and van Campo, 1994; Wei and Gasse, 1999). This suite of palaeoclimatic data leaves little doubt that the whole South Asian monsoonal system was strengthened during early Holocene times. Similar pollen and lake-level sequences show wetter early to mid-Holocene climates in some other parts of tropics, such as the Caribbean and northern Australia (Hodell et al., 1991; Shulmeister, 1992).

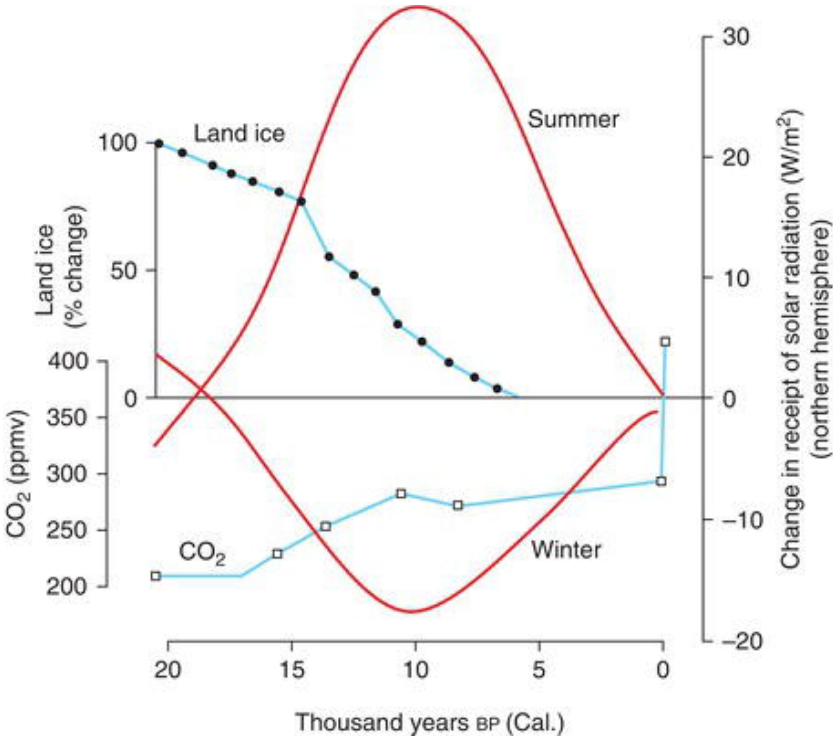
Increased rainfall may also have been partly responsible for the dominance of sub-humid forest in the wetter parts of the Mediterranean basin during the first half of the Holocene. Enhanced early and mid-Holocene precipitation is recorded in a range of other natural archives from the central and eastern Mediterranean region, including stable isotopes from cave speleothems (Bar-Matthews et al., 1997) and lakes (Roberts et al., 2008). The same is true of the Mediterranean-type biome elsewhere in the world, such as in South Australia, in the form of higher-level freshwater lakes and – significantly – from grass pollen indicative of enhanced summer monsoonal rainfall (Bowler, Singh in Chappell and Grindrod, 1983; Luly, 1993). Increases in rainfall may have been less marked here than in the tropics (+20% has been suggested in South Australia, and 25–40% in the eastern Mediterranean; Jones et al., 2007), but on account of their seasonal distribution, they had a potentially critical effect on plant ecology.

Shifts in atmospheric circulation during the early and mid-Holocene did not bring increased rainfall to all mid-latitude regions. Those areas out of reach of sub-tropical moisture sources instead became drier than they are at present, for example, on the North American prairie between the Rocky Mountains and the Mississippi. The mid-west prairie–forest boundary moved eastwards until 8000 Cal. yr BP but subsequently shifted back in the opposite direction (Webb et al.,

1984). As this boundary is today controlled by moisture availability, it is reasonable to use it as a proxy climatic indicator, recording climatic conditions between 8 and 4.5 ka BP which were drier than at present. This climatic phase is also recorded in sediment cores from the many small lakes of the North American Great Plains region (e.g. Grimm et al., 2011). Diatom and geochemical analyses from Devil's Lake in North Dakota, for example, show high salinities during the first half of the Holocene, while before and after this phase the lake water was relatively fresh (Fritz et al., 1991; Haskell et al., 1996).

A sequence of vegetation reversion is also evident further north in America at the boundary of boreal forest and tundra (Ritchie, 1987). Between 13 ka and 8 ka BP, Canada's boreal spruce forest moved northwards, only to retreat southwards after 4.5 ka BP (see Figure 4.7). In the highlands of New Guinea and South America too, vegetation belts lay 100–200 m above modern elevations between c.10 ka BP and the mid- to late Holocene (Flenley, 1979b). Climatically calibrated pollen data for Europe similarly suggest that most of Europe had higher temperatures than at present (Huntley and Prentice, 1988; Cheddadi et al., 1997; Davis et al., 2003). However, some regions like the Mediterranean basin and parts of the tropics may have been slightly cooler, linked to increased cloud cover and soil moisture, so that Iversen's idea of a mid-Holocene 'thermal optimum' may not have world-wide applicability (Jansen et al., 2007).

Figure 4.10 Changing receipt of solar radiation and other climatic controls since the LGM (adapted from Kutzbach and Street-Perrott, 1985).



The cause of these climatic changes lies in astronomically controlled variations in the Earth’s receipt of solar radiation (see [Figure 4.10](#)). At the start of the Holocene, one of the three Milankovitch cycles – namely, the precession of the equinoxes – lay at the opposite point on the cycle from that of the present day. This meant that the northern hemisphere then received nearly 8% more solar radiation during summer months than it does at present. Heating over the land, especially South Asia, caused low-pressure conditions which sucked in humid, monsoonal air masses from the south. Numerical models of the Earth’s atmospheric circulation simulate this strengthened monsoon and an increase in effective precipitation over northern hemisphere land areas (see [Table 4.3](#)). These models also show good agreement with Holocene palaeoclimatic data for most of Africa and South Asia (Kutzbach and Street-Perrott, 1985; COHMAP Members, 1988; Wright et al., 1993; Braconnet et al., 2004). The climate of this early and mid-Holocene ‘thermal optimum’ has been used to test and calibrate numerical climate models. These are the same computer models that are used to simulate the effects of the current enhanced ‘greenhouse’ effect on global climate during the twenty-first century (Joussaume et al., 1999). However, the early Holocene does not offer a direct analogue for a future greenhouse gas-warmed climate, because it was dominated by seasonal, rather than mean annual, changes in the Earth’s net receipt of solar radiation (Mitchell, 1990; Jansen et al., 2007).

Table 4.3 Mean annual hydrological changes between 9 ka BP and the present day

	Global		Northern hemisphere only	
	Land (%)	Ocean (%)	Land (%)	Ocean (%)
Evaporation (E)	+1.3	−2.4	+2.1	−2.6
Precipitation (P)	+9.7	−6.0	+12.0	−4.0
P−E	+8.5	−3.6	+9.9	−1.4

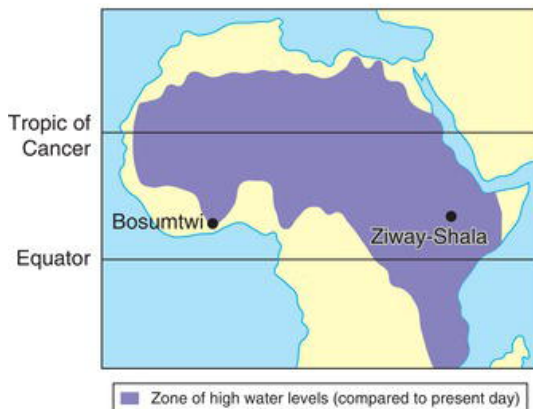
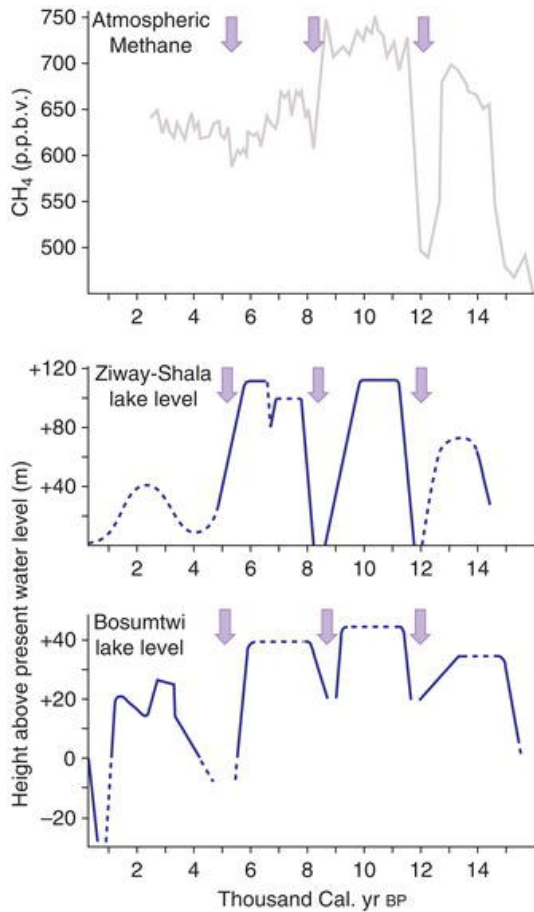
Positive values record an increase at 9 kaBP compared to the present

Source: Kutzbach and Gallimore (1988). Reproduced with permission of the American Geophysical Union.

In addition to direct changes in the receipt of solar radiation, orbital ‘forcing’ of Holocene climate appears to have been amplified by other parts of the Earth system, such as the biosphere. Modelling experiments have shown that vegetation has had a significant effect on climatic conditions in key boundary areas, such as the edges of forest and desert biomes, by changing the surface albedo and moisture conditions. Along the boreal-tundra ecotone, for example, the expanded forest has been calculated to have itself caused an additional warming of about 4°C in spring temperatures 6000–7000 years ago, over and above that caused directly by solar heating (Foley et al., 1994). In comparable fashion, the Saharan and Arabian deserts shrank by much more than predicted directly by the models during the first half of the Holocene because of feedback effects (Street-Perrott et al., 1991; Kutzbach et al., 1996). It seems that once the desert was transformed to savanna, it had an inherent tendency to stay that way. Even with land cover and

ocean feedbacks amplifying climate change, computer models have only been able to simulate a greening of the Sahara as far as about 20°N (Braconnet et al., 2004). North of this, the models predict that desert conditions were maintained when field evidence tells us that most of the landscape was under Sahelian scrub or steppe vegetation.

Figure 4.11 Tropical African lake-level records compared to atmospheric methane concentration recorded in a Greenland ice core (data from Gillespie et al., 1983; Talbot and Delibrias, 1977; Blunier et al., 1995). Arrows show abrupt arid events interrupting the longer-term climatic trends. The map shows the extent of the zone of high lake levels during the early Holocene African Humid Period.



Holocene climatic changes also offer a warning about the hazards of climate prediction. On several occasions during the Holocene, the climate deteriorated

abruptly, only to return to its previous state within a few centuries or less (see [Figure 4.11](#)). These climatic crises are recorded in abrupt falls in the levels of many of Africa's lakes dated to c.12 ka, 8.2 ka and 5.2 ka BP (Street-Perrott and Roberts, 1983; Gasse, 2000). As with the rapid rises in temperature at the Pleistocene–Holocene transition, these hydrological changes sometimes occurred so rapidly that their true rate cannot be determined by ^{14}C dating. The ecological consequences of these arid episodes were sometimes catastrophic; for instance, at Lake Shala in Ethiopia, one lake recession resulted in the death of fish on a massive scale (Gillespie et al., 1983). On the other hand, pollen analysis indicates that trees – being long lived – were, in some cases, able to survive through the extended droughts (Lamb et al., 1995). The abrupt mega-drought events which interrupted the African Humid Period have their equivalent in the Greenland ice core record. This shows sharp climatic cooling associated with drops in atmospheric methane concentration (Thomas et al., 2007), implying a reduction in tropical wetlands which are a major source of this 'marsh gas' (see [Figure 4.11](#)). These mega-droughts resulted from short-lived mechanisms such as the switching off of the Atlantic Ocean 'conveyor' (see Chapter 3) rather than long-term orbital changes in solar radiation. The short-lived but intense '8.2 ka' (8200 Cal. yr BP) event, for example, appears to have been linked to the final catastrophic demise of the Laurentide ice sheet over Hudson Bay at about the same time, which led to a cold meltwater pulse into the North Atlantic (Barber et al., 1999; Alley and Ágústsdóttir, 2005). The 8.2 ka event – which provided a (very) loose basis for the disaster movie *The Day After Tomorrow* – seriously disturbed the global climate for about a century, led to failure of monsoon rains in China and raised sea levels by around 2 m (Morrill and Jacobsen, 2005; Daley et al., 2011; Törnqvist and Hijma, 2012; Liu et al., 2013). Not surprisingly, abrupt events such as this create considerable uncertainty for predicting future climates.

Conclusion

The key theme of the early part of the Holocene is one of adjustment and adaptation to new environmental conditions. Sea level adjusted to the reduced global volume of ice and in turn modified river base levels and coastal ecologies. Soil–vegetation associations responded to climatic warming by dramatically changing their extent and distribution. The world's deserts and tundra became much less extensive than they had been under a glacial regime, while temperate grasslands expanded, at least during the first half of the Holocene. Above all, forest ecosystems – both temperate and tropical – were greatly enlarged in extent as the higher post-glacial temperatures and moisture encouraged tree growth, so increasing global net primary productivity (NPP) and biomass (see [Table 3.3](#)). All these ecosystems brought with them a host of insects, birds and mammals. On the expanding North American prairie, for example, bison were presented with a new and largely vacant ecological niche in which they were able to increase their populations enormously. But biomes did not move from their glacial to their interglacial positions *en masse* like a well-disciplined army. Rather, different

species set off from different places and arrived at different times, like the anarchic race at the Mad Hatter's tea party in *Alice in Wonderland*. In other words, migrations and adaptations were achieved by individual taxa rather than by whole plant formations.

It was not only physical and biological regimes that had to adapt; so too did cultural ones. The advent of the Holocene brought radically altered environmental resources to the world's human population. Mammalian extinctions and the reduction in open habitats such as tundra meant a reduced dependence on meat and other hunted animal products. There was also habitat loss in the coastal zone as sea levels rose. By contrast, forests and wetlands increased, and it is no surprise that from the European Mesolithic to the Saharan Aqualithic, early Holocene habitations are found on what were then lake shores (Sauer, 1948). It was believed by classical authors such as Varro and Seneca that there had once been a 'Golden Age', '...when man lived on those things which the virgin earth produced spontaneously...' and when '...the very soil was more fertile and productive' (cited in Glacken, 1967, p. 133). If ever there was such a 'Golden Age', then it was in the early Holocene when soils were still unweathered and uneroded and when Mesolithic peoples lived off the fruits of the land without the physical toil of grinding labour (Sahlins, 1974). Human life in the Mesolithic was adjusted to the movement of deer or salmon; the autumn harvest of fruits, nuts and berries; and other rhythms of nature in the same way it was in h-f-g groups like the Kwakiutl Indians of America's northwest Pacific coast, who were encountered by Europeans in the late eighteenth and nineteenth centuries.

On the other hand, even in the early Holocene, the human role in nature was far from passive. Although there was little long-term benefit to be gained from wholesale cutting down of trees – as it was the forest that provided the resources necessary for human life – human groups were well capable of manipulating those plant and animal resources of use to them. There is an unusually high concentration of ivy (*Hedera helix*) pollen at many European Mesolithic sites, for example; the most likely explanation is that ivy was used to attract red deer (Simmons and Dumbleby, 1974). Similarly, hazel, which coppices freely in response to burning, is much more abundant in the protocratic phase of the Holocene than in any previous Pleistocene interglacial – possibly an indirect result of Mesolithic use of fire (Smith, 1970; Huntley, 1993b). The use of fire and other human impacts meant that the 'wildwood' of the early and mid-Holocene may not have been in quite the 'virginal' state Varro imagined.

Even more portentous were changes that were taking place in the subsistence base. Although the dominant mode of production remained founded in seasonally mobile h-f-g economies, the food base became broader and more diversified. In a few environments, such as the basin of Mexico, it proved possible to schedule hunting and collecting of seasonal resources in such a way that foodstuffs were available in one place all year-round (Niederberger, 1979). A related form of early Holocene adaptation involved intensive use of certain non-tree food plants, particularly grasses. The combination of harvesting wild grass seeds and a sedentary existence was to lead to an economy fundamentally different from all that had gone before, one that was to irreversibly change human relations with nature. That economy was farming.

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The first farmers

Agricultural origins

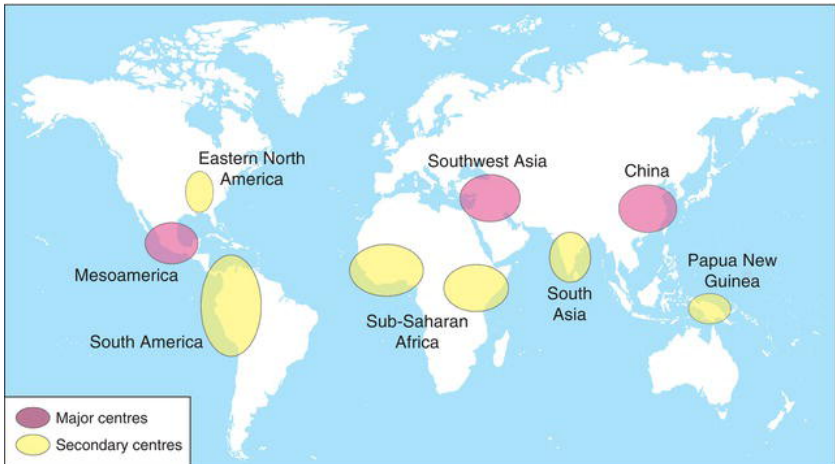
In the previous two chapters, attention was focused on the transition from a glacial to an interglacial world and the consequences of climatic change for natural and cultural ecologies. In this chapter, we turn to humans as agents of environmental modification, more specifically to the advent of agricultural modes of production. These in turn were made possible by the domestication of plants and animals.

The question of where and when our domestic crops and animals originated is a truly interdisciplinary problem which has received contributions from – amongst others – geographers such as Carl Sauer (1952), archaeologists such as Eric Higgs (1972) and David Harris (Harris and Hillman, 1989; Harris, 1996a) and crop scientists such as Jack Harlan (1975, 1995). However, our story starts with the Russian plant geneticist Nikolai Vavilov who in 1926 proposed eight hearth areas as the most likely centres for the origin of individual crop complexes (Harris, 1990). Most of these lay in a belt across the southern sub-tropical margins of Asia and Europe, the others being in Central and South America and Ethiopia. Vavilov based his hearths on areas of maximum genetic diversity which he thought would include the nearest wild relatives to plant domesticates. This assumed that the plant distributions have not changed significantly since the time of their domestication, for instance, as a result of climatic change. Also underlying Vavilov's work was the belief that domestication of particular forms occurred only at specific times and places, rather than being a continuing process that was diffuse in time and space. The present, much wider distribution of domestic crops was thought to be a result of subsequent diffusion from the original hearths into adjacent cultures.

Subsequent genetic studies of living plants (e.g. Jones and Brown, 2000) have narrowed Vavilov's possible locations for plant domestication, but on their own, they cannot confirm if it occurred where he suggested, nor when. This dearth of historical information has been countered since the mid-twentieth century by collaboration between archaeologists on the one hand and botanists and zoologists on the other. The latter have studied the animal bones and charred or waterlogged seed remains from excavated archaeological sites. Integrated bio-

archaeological research has produced some of its most successful results in southwest Asia, Mesoamerica and China (Figure 5.1). In line with Vavilov's predictions, these regions have turned out to be the homelands of three of the world's most important food crops, namely wheat/barley, maize and rice.

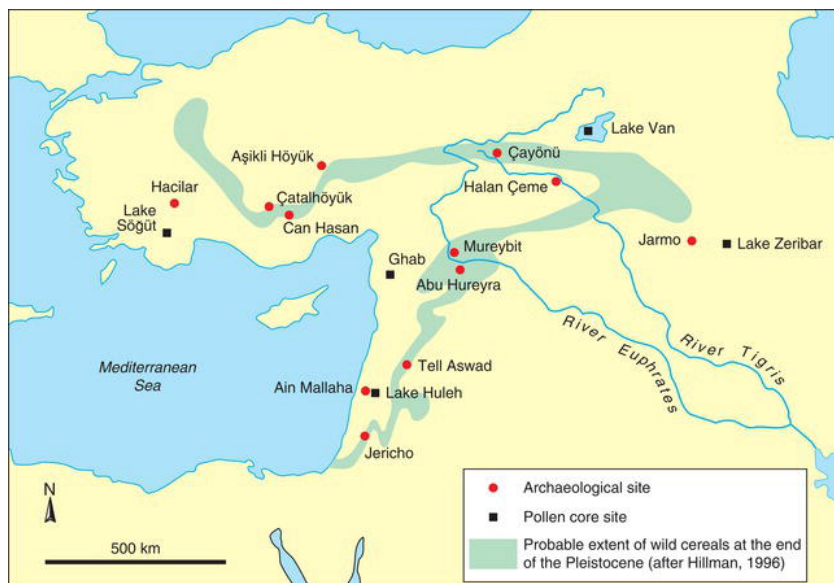
Figure 5.1 World centres of plant and animal domestication.



Southwest Asia

Southwest Asia (alternatively known as the Near East) is probably the earliest and arguably the most important of all the global centres of domestication. From the Neolithic farmers of this region, we have inherited cereal crops such as wheat (*Triticum*), barley (*Hordeum*) and rye (*Secale*), pulses including pea (*Pisum*) and lentil (*Lens*) and a range of domestic animals including cattle (*Bos*), sheep (*Ovis*), goat (*Capra*), pig (*Sus*) and dog (*Canis*). The increased economic productivity and new social organisation produced by this 'Neolithic revolution' at the beginning of the Holocene was associated with population increase and great developments in material culture.

Figure 5.2 The southwest Asian centre of domestication.



Integrated bio-archaeological field research into the origins of agriculture was initiated in the 1950s by American Robert Braidwood among others. Excavations of mounds – or ‘tells’ – such as Jarmo and Jericho revealed Neolithic villages ^{14}C dated back to the very start of the Holocene (Figure 5.2). Their mudbrick dwellings contained charred seeds, whose subsequent analysis, notably by Danish archaeobotanist Hans Helbaek, showed many of them to be early domestic cereals significantly different from naturally occurring wild varieties (Technical Box VI). A big advantage of this work is that it could be compared against a rich modern ethnobotanical record, for traditional agriculture remains little changed in some parts of the southwest Asia from that which existed during Neolithic times (Hillman, 1984) (see Plate 5.1). The identification of domestic varieties was less easy with animal bones, but criteria such as size and age–sex structure of the animal population indicated that these were herded rather than hunted (Meadow, 1989; Clutton-Brock, 1999). The success of this work in turn encouraged further research, and in consequence, there is now a relatively detailed record of the cultural and economic prehistory of the southwest Asia between about 15 and 8 ka BP (i.e. terminal Pleistocene and earliest Holocene) based on bio-archaeological data. Although some areas such as Afghanistan are still poorly known, enough information now exists to suggest with some confidence when and where cereal-based farming began (Zohary and Hopf, 2001).

Plate 5.1 Traditional agriculture in Anatolia: this threshing sled used flints to separate kernels from their husks.



Archaeological sites older than c.12 ka BP (12 000 Cal. yr BP) belong to final (or Epi-) Palaeolithic-type cultures, such as the Natufian culture of Palestine (Bar-Yosef and Belfer-Cohen, 1992). Sites were relatively small and most were only occupied seasonally. Many site economies included harvesting of wild cereal stands, while others involved husbandry of essentially wild herds of sheep, goat and gazelle – but in both cases, these formed part of broadly based economies utilising a wide range of resources (Hillman, 1996; Harris, 1998; Rosen and Rivera-Collazo, 2012). During the earliest pre-pottery Neolithic (11.7–10.5 ka BP), there are clear suggestions of experimental or proto-agriculture, with morphologically wild cereals found outside their natural habitats, notably in high water-table alluvial environments such as the Euphrates floodplain at Mureybat (Sherratt, 1980). By the later pre-pottery Neolithic (10.5–9 ka BP), fully fledged farming villages appeared widely throughout the so-called Fertile Crescent and indeed beyond it into central Turkey. Sites were much larger, reflecting communities whose populations were numbered in hundreds or even thousands and whose mode of production was rapidly becoming based on agriculture.

TECHNICAL BOX

Technical Box VI: Morphological changes in plant domesticates

Under natural conditions the seeds of cereal grasses such as wild einkorn (*T. boeoticum*, [Plate 5.2](#)) would have been attached to a rachis or central stem which shattered when the seeds were ripe, thus aiding seed

dispersal. Once harvested by sickle, on the other hand, a tough rather than a brittle rachis became advantageous. Regular harvesting by sickle therefore eventually led to the selection of mutant strains whose seed heads did not shatter when ripe. In similar fashion, threshing of the seeds tended to select strains whose grain separated from the chaff, and these became free-threshing wheats. Awns, the hair-like tails which help to bury wild wheat seeds in the soil, so protecting them from birds or from fires (Elbaum et al., 2007), are also often lost with domestication (Fuller and Allarby, 2009). Most of the first wave of plants to be domesticated were self-pollinating (Zohary and Hopf, 2001). This isolated them reproductively from wild progenitors and meant that farmers could grow desirable cultigens in the same area as wild varieties without the former being genetically ‘swamped’.

Plate 5.2 Wild einkorn (*Triticum boeoticum*) from southwest Asia: one of the main ancestors of modern wheat.



Human pressures such as these eventually led to changes in the plant that made it a distinctively different, domestic species. Domestic wheats have more chromosomes, larger seed heads (and therefore greater yields), simultaneous ripening and other features (Fuller, 2007); but they are also dependent on human beings for propagation, and hence survival. In fact, a main diagnostic feature of domesticated plants is the loss of the ability to disperse seed without human assistance (Blumler and Byrne, 1991). Domestication can therefore only be recognised from

plant morphology some time after the selective processes began, because of the time lag required for species' genetic make-up to be modified. Experimental harvesting of morphologically wild cereal stands growing has shown that this time lag could potentially have been short (less than 200 years) even without deliberate selection, but only once certain key harvesting methods were adopted (Hillman and Davies, 1990). Harvesting of a crop by beating into baskets, for example, would leave the seed crop remaining in its morphologically wild 'shattering' form.

Sites of domestication for founder crops were therefore restricted to the nuclear area of the Fertile Crescent (Nesbitt, 2002). There is good bio-molecular and archaeobotanical evidence that single-grained einkorn wheat was domesticated in the Karaca Dağ volcanic uplands of southeast Turkey during the first 1500 years of the Holocene (Heun et al., 1997; Willcox, et al., 2009). Morphologically domestic emmer wheat appears early in the Damascus basin and also further north near the middle reaches of the river Tigris at Çayönü, suggesting the possibility of multiple independent domestication events, and the same is true for barley (Garrard, 1999; Colledge et al., 2004). The first domestic sheep and goat appear at about the same time as these large-grained cereal crops (Conolly et al., 2011), with the earliest dated bio-archaeological evidence coming from the Zagros mountains of western Iran and northern Iraq (Zeder and Hesse, 2000).

Evidence therefore shows that the change from a mobile hunter-gatherer economy to sedentary agriculture in the southwest Asia was not instantaneous, but involved a transitional stage of experimentation, partial dependence on domesticates and variability in individual site economies (Zeder, 2011). Only when the new systems of food procurement had proved their worth were the various elements brought together and adopted as a single economic 'package'. Some researchers (e.g. Higgs, 1972) have aimed to demonstrate that the earliest evidence of domestication for individual species in the southwest Asia lies not in the Neolithic but in the preceding final Palaeolithic period. Though significant in itself, this should not cause us to forget that fully agricultural economies only emerged in later pre-pottery Neolithic times, around 10 000 years ago. It was the adoption of this Neolithic agricultural package rather than its original creation that occurred quickly, typically within a few hundred years. At Abu Hureyra, for example, there was an abrupt and permanent shift from gazelle (>80% of all animal bones) to sheep/goat (c.70%) between early and late pre-pottery Neolithic levels (Legge, 1996). One of the reasons for the success of the cereal-based Neolithic economy was its incorporation of animal domesticates to form an integrated agricultural package. This package of livestock and grain crops was

subsequently ‘exported’ from the southwest Asia into adjacent areas, especially northwestwards to form the basis of traditional European peasant farming.

China and South Asia

The second of the world’s great cereal crops is rice, of which the most important type is that indigenous to East and South Asia (*Oryza sativa*). Much important bio-archaeological field research has taken place here in recent years, and a coherent story for the origins of Asia’s rice cultures is now emerging. The ancestor of *O. sativa* was the wild rice *Oryza rufipogon*, and the wild relatives of domestic rice can be found over much of southeast Asia, even in Australia (see [Figure 5.3](#)), although Chang (1976) proposed a more restricted area of origin for cultivated rice, in an arc stretching from northeast India through Burma, Thailand and Vietnam to southern China. During the terminal Pleistocene and earliest Holocene, much of this area of southeast Asia lay within the province of the Hoabinhian archaeological culture. A wide range of plant and animal resources were exploited at Hoabinhian sites such as Spirit Cave in Thailand, but there are few – if any – signs of incipient rice cultivation here. Even at the much later Thai sites of Ban Chiang and Non Nok Tha (c.6 ka BP onwards), carbonised rice glumes that were preserved may not be morphologically domestic. This suggests that early manipulation of wild rice took place at the edges of its natural range, and bio-archaeological evidence shows that it was in China’s Yangtze lowlands, rather than in southeast Asia, that rice’s true potential as a cultigen came to be realised.

The earliest known communities of the Yangtze River valley which included rice as a significant foodstuff were those of the early Neolithic, dating back to around 8.5 ka BP (Glover and Higham, 1996; Cohen, 1998). Much of the evidence comes from microscopic rice phytoliths rather than from rice grains proper. However, at the site of Hemudu dating from approximately 7 ka BP, waterlogged conditions led to the preservation of a layer of stalks, grains and husks of rice spread over 400 m² and up to half a metre thick, plausibly an ancient threshing floor. Both morphologically wild and domestic rice were found at the site, as were hoes made of ox shoulder blades and bones of domesticated water buffalo, an animal long associated with lowland rice culture. This combination suggests that farming was based on fixed field horticulture rather than a shifting form of cultivation. Shifting cultivation may have been more appropriate in the drier loess lands to the northwest where the contemporary Yangshao villages depended on foxtail millet (*Setaria*) and broomcorn millet (*Panicum*) rather than on rice (An, 1989). These millet cultures seem to have developed in parallel with the rice cultures further south, exploiting the cooler and drier climatic and environmental conditions of northern China.

Chinese Neolithic sites include a range of other domesticates, such as pig, chickens and water chestnuts (actually not a nut at all but the corm of a marsh sedge), but wild species, such as turtle, deer, acorns and other true nuts, continued to be significant (Fuller et al., 2008). In other words, foods deriving from hunting, fishing and foraging do not seem to have been displaced as completely as they were in southwest Asia. Nor did economies based on rice

cultivation spread as rapidly into adjacent regions of East Asia as their equivalent cereal cultivators across Europe. In Japan, for example, the Neolithic Jomon culture had established a successful economy based on fishing and plant collecting during the early Holocene, and it seems to have been distinctly reluctant to adopt crop plants from the Asian mainland. Although people would have known of the ‘advantages’ of grain cultivation for several millennia, rice cultivation only became significant in Japan around 3 ka BP (Bellwood, 2005). Similarly, the earliest evidence for domesticated, non-shattering rice on the mainland and islands of southeast Asia occurs 2000 to 4000 years later than in China’s Yangtze valley, indicating a rather gradual uptake of farming southwards.

The Indian sub-continent occupies a place that is intermediate between the great crop hearth areas to its east and west; even today rice and naan bread are offered as a choice to accompany dishes in Indian cuisine. Wheat, barley, sheep and the rest of the southwest Asian agricultural ‘package’ form the basis of farm economies in northern India and the Indus Valley, while rice cultivation dominates the agrarian landscape in South India and along the river Ganges. It is a matter of debate how far this pattern resulted from simple spread into South Asia from adjacent regions and how far there was independent domestication of crops and livestock. India forms part of the natural range of wild rice (Figure 5.3), and it is conceivable that the long-grained variety of cultivated rice, *O. sativa var. indica*, was domesticated locally from the wild progenitor. However, genetic and bio-archaeological evidence favours a single origin for domestic rice, namely, in China during the early Holocene, with the Indian variety splitting off from already domesticated short-grained rice in the vicinity of the Ganges valley about 4000 years ago (Molina et al., 2011). Dorian Fuller (2006) has argued that there was indigenous domestication in South Asia, for example, of some millets and pulses in India’s southern Deccan region and of zebu cattle in the northern part of the subcontinent. However, at a global scale, it is clear that the Indian sub-continent was not a major independent centre of domestication as Vavilov had envisaged.

Figure 5.3 The south and east Asian centres of domestication.



Mesoamerica

The major cereal crop of New World origin is maize (*Zea mays*, also called corn, as in ‘corn flakes’), whose probable place of origin is central (or Meso)america. For many years, P.C. Mangelsdorf (1974) argued that the wild ancestor of maize was no longer to be found, having been made extinct, just as wild cattle (or aurochs, *Bos primigenius*) have been in Eurasia. It is now believed, however, that ancestral maize was probably descended from teosinte, a wild relative of maize still found in Mesoamerica (Wilkes, 1989; Martienssen, 1997). Teosinte has a small and fragile ear (cob), seeds embedded in the rachis and hard glumes – in other words, similar features to those that differentiate wild from domestic cereals in southwest Asia and China.

Projects at Tehuacán and Oaxaca in southeast Mexico have traced the Holocene evolution of maize cultivation in archaeobotanical records, with tiny maize cobs less than 3 cm long recovered from archaeological contexts of the Archaic period. The oldest domestic maize cobs dated directly by AMS ^{14}C go back to approximately 6.5 ka BP (Piperno and Flannery, 2001), but supplementary evidence such as maize pollen, phytoliths and starch grains indicates that the technical knowledge and potential cultigens needed for a farming economy were in existence before this. Some of the oldest bio-archaeological evidence comes from the Rio Balsas region southwest of Mexico City, where maize appears to have been cultivated soon after 9 ka BP (Piperno et al., 2009). The fact that these remains were found in the proposed homeland of teosinte further supports the

idea that this region was a probable locus of maize domestication. Despite this, and even more markedly than in East Asia, there appears to have been no rush in the Americas to abandon hunter-foraging and take up farming. Garden cultivation of maize and other plant domesticates such as squash (*Cucurbita*), chili pepper (*Capsicum*) and avocado (*Persea*) were simply added to existing forms of food procurement (Pickersgill, 1989; Smith, 1997). Through selective pressures, maize cobs became progressively enlarged over time (see [Plate 5.3](#)), and a critical threshold may have been reached around 4.3 ka BP when, with yields of 200 kg/ha, maize became more productive than any wild plant resource. Only at this point, it is suggested, was maize-based farming capable of supporting settled farming communities (Bellwood, 2005), and it is at this date that fully developed agriculture first appears in the New World (see [Figure 5.4](#)). A potentially different story has been suggested from the measurement of the stable isotopes of carbon. Isotopic analyses of human skeletons from Tehuacán suggest that cereal grains already formed a major component of diet as early as 7 ka BP (Farnsworth et al., 1985). This apparent discrepancy in evidence may have been due to intensive harvesting of grasses such as *Setaria* millet, which have the same isotopic signature as maize. In other parts of Mexico, such as the Gulf coast of Tabasco, forest clearance for maize cultivation is also already evident by this time (Pope et al., 2001), even if wild resources remained a major component of diet until the last four millennia (Blake et al., 1992).

How does the pattern of domestication in Mesoamerica compare with those in southwest Asia and in China? Similarities include a recognisable nuclear hearth area within which agriculture originated, and early Holocene dates for the beginning of this process. On the other hand, the differences between the three regions are no less striking. Whereas agriculture quickly replaced hunter-foraging as the main source of food in southwest Asia, in Mesoamerica, this process took at least 3500 years, with East Asia somewhere in between. One reason for this delayed reaction may have been the paucity of animal domesticates in the New World, in turn partly due to the massive loss of large mammal species at the end of the last Ice Age, including potential domesticates such as the wild horse. The native American llama (*Lama*) and guinea pig (*Cavia*) are relatively unimportant food resources, and they were not integrated with crop husbandry as cattle came to be in the Old World for traction and manure. A further difference is that southwest Asian domestication was essentially a once-and-for-all occurrence squeezed into two or three millennia immediately after the end of the Pleistocene, but in the New World, it has been a continuing process throughout the Holocene (see [Figure 5.5](#)).

Plate 5.3 Domesticated maize (corn) cobs from archaeological excavations at Tehuacán, Mexico. The oldest one (c.7500 Cal. yr BP), at only 3 cm long, is hardly larger than its wild ancestor, while the most recent (500 years old) is similar to modern maize. © Robert S. Peabody Museum of Archaeology, Phillips Academy, Andover, Massachusetts. All Rights Reserved. Reproduced with permission of the Robert S. Peabody Museum of Archaeology.

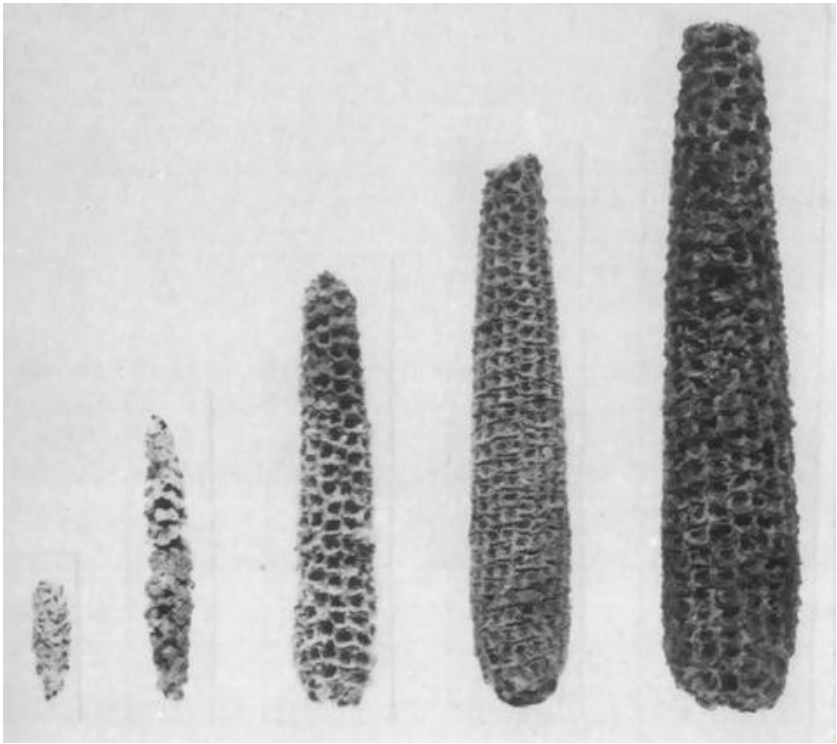
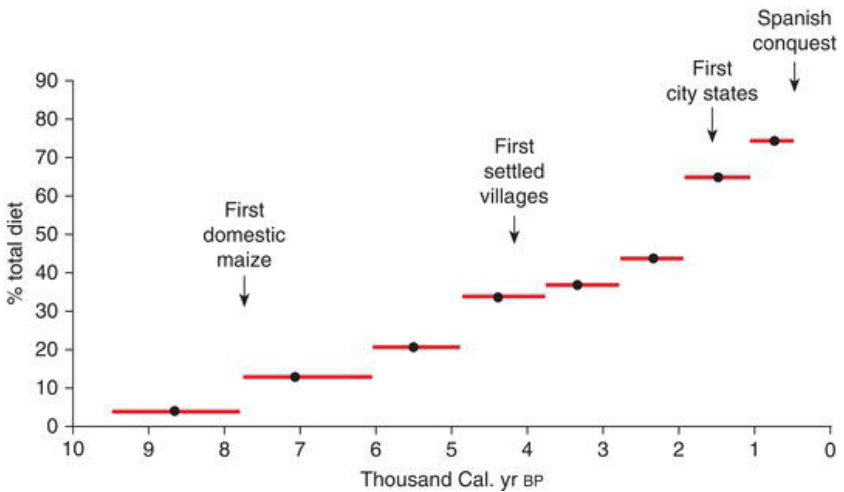


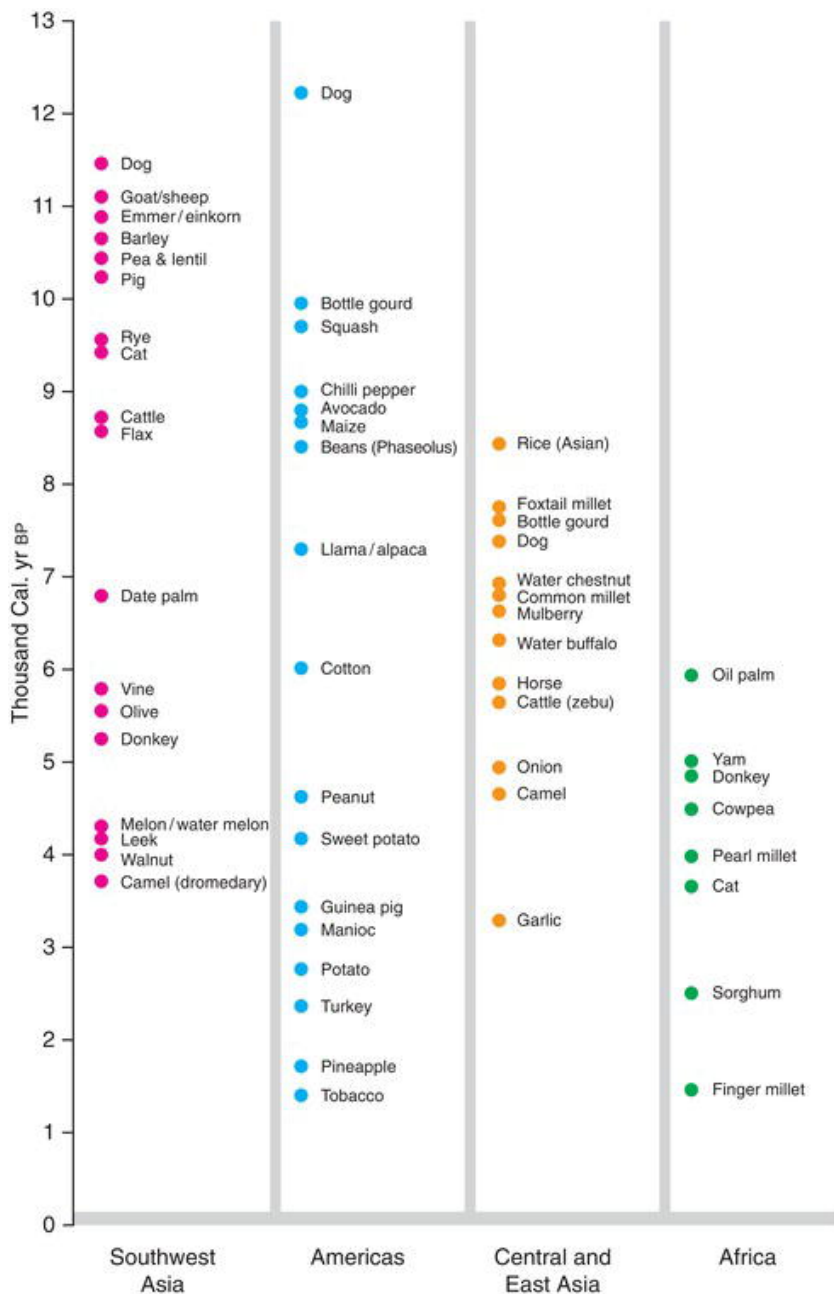
Figure 5.4 The changing contribution of domestic plants to total diet, Tehuacan, Mexico (data from MacNeish, 1967).



The ongoing nature of the latter has much to do with the fact that New World domestication was not restricted to Mesoamerica, even if its principal focus was there (see [Figure 5.1](#)). Some crops, such as the sunflower (*Helianthus*), originated

north of Mexico, most importantly in eastern North America (Smith, 2006). While there is evidence of independent plant domestication along the middle reaches of the Mississippi and Ohio rivers, this region – like South Asia – did not evolve to become a globally significant player in terms of the origins of world agriculture. Much more important have been crops derived from South America, both of highland and lowland origin. Because teosinte is not found south of Guatemala and because maize appeared in an already developed form in South America around 6 ka BP (Bush et al., 1989), it has often been believed that maize was exported southwards from Mesoamerica rather than being an independent South American innovation. The genetic evidence also tends to support a single rather than multiple origins for maize domestication (Matsuoka et al., 2002). On the other hand, the majority of maize varieties found in Peru are endemic to that country, suggesting that there were secondary centres of maize hybridization in South America (Wilkes, 1989). Maize apart, however, the history of New World domestication is very much one of continuing interaction and exchange of species between different regional centres. Carl Sauer characterized this in terms of a Mesoamerican origin for seed agriculture and a South American origin for vegetatively reproducing plants such as manioc. Although this distinction appears broadly valid, there is evidence in favour of multiple domestication of at least some plants; for instance, beans (*Phaseolus*) appear early and independently as domestic crops in both Mexico and Peru (Harlan, 1975; Chapter12).

Figure 5.5 Recorded first appearances of individual domesticates in different regions during the Holocene.

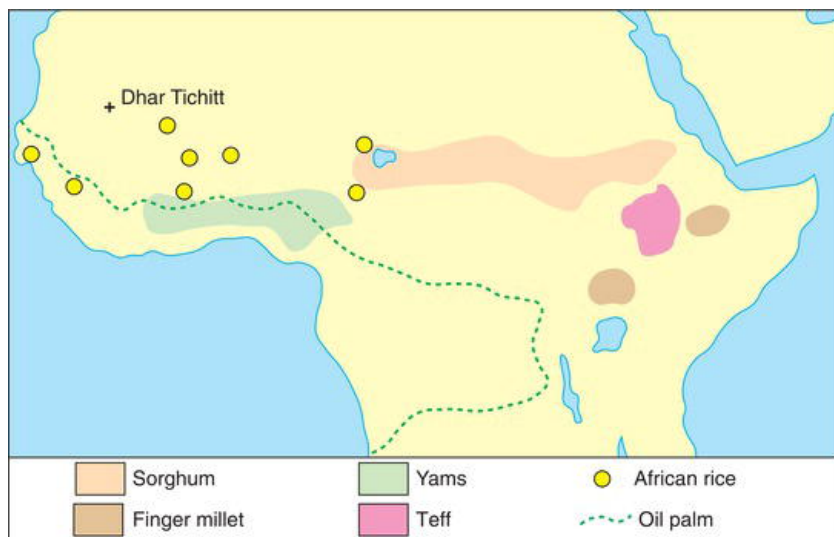


Tropical domesticates

The tropics have long been the ‘cinderella’ regions of plant and animal domestication. This has been explained in terms of the poor preservation of root and tuber crops in archaeological sites, compared with seed crops such as wheat and maize. Furthermore, asexual reproduction inhibits genetic change and makes cultural modification of vegetatively reproducing plants harder to detect (Hawkes, 1989; Hather, 1994). The dates at which crops such as manioc (*Manihot*), yams (*Dioscorea*) and taro (*Xanthosoma*) were domesticated are consequently rather sketchy, although all are of undoubted tropical origin. On the other hand, the magnitude of this problem has often been exaggerated. There are, for example, tropical environments such as the Peruvian desert in which plant remains are exceptionally well preserved. Even more significant is the importance of tree, fibre and seed as well as root-tuber crops in the tropics; there are, in fact, more tropical cereal crops than there are mid-latitude ones!

This is well illustrated in the case of sub-Saharan Africa, where nine wild indigenous grasses are known to have been domesticated (Shaw, 1977). One of the most important of these is sorghum or guinea corn (*Sorghum bicolor*) whose different domestic races exhibit clear regional patterns across the continent. From these, Jack Harlan and Ann Stemler (1976) postulated an evolutionary sequence starting in the eastern Sudanian zone (see [Figure 5.6](#)), with later secondary centres in the western Sudanian zone, east central Africa and also in northwest India. Testing this hypothesized sequence bio-archaeologically should be no more difficult than researching the origins of wheat, but despite claims of early domestication in the Sudan, firm archaeobotanical evidence for domestic sorghum is little more than 2000 years old (Wetterstrom, 1998). One of the few cases where early adoption of an African crop has been demonstrated is at Dhar Tichitt in the western Sahel, where an initial reliance on harvesting of wild grasses was replaced around 3200 years ago by cultivation of pearl or bulrush millet (*Pennisetum*). The paucity of bio-archaeological evidence notwithstanding, likely areas of domestication for native African crops have been proposed (see [Figure 5.6](#)), and they suggest a much more spatially diffuse pattern than in the Mesoamerica, China or southwest Asia.

Figure 5.6 Probable area of origin of African plant domesticates (data from Shaw, 1980, and Harlan and Stemler, 1976).



Clear centres of domestication of the type proposed by Vavilov do not seem to be typical in low latitudes generally. Thus, in South America, manioc is a crop of lowland origin – perhaps from the Orinoco basin – while the potato (*Solanum*) is native to the Andean mountains (Pickersgill and Heiser, 1977; Hawkes, 1989). However, any temptation to regard tropical agriculture as a chronologically late offshoot of mid-latitude cereal cultures should be resisted. Domestic crops appear as early as 9000 years ago in South America (Piperno and Pearsell, 1998; Piperno and Stothert, 2003), while the first farmers on the south side of the Sahara grew locally domesticated crops, not imported ones. But the most dramatic – and unexpected – evidence for early independent agriculture in the tropics comes from the highlands of Papua New Guinea. Agricultural developments involving both swamp and dryland cultivation of yams, taro, sugar cane and bananas, probably along with pig husbandry, occurred early in the Holocene (Denham et al., 2004). Some of the earliest records come from excavations of field drainage systems at Kuk Swamp (Bayliss-Smith, 1996; Denham et al., 2003; Haberle et al., 2012). These date back to 6.8 ka BP with pollen and charcoal indicative of forest burning and clearance and geomorphological evidence of swamp disturbance as early as 10 ka BP. This potentially puts Papua New Guinea within a whisker of having the oldest known agriculture anywhere.

On the other hand, unlike southwest Asia or Mesoamerica, Papua New Guinea does not seem to have acted as a major centre of diffusion for its newly acquired plant and animal domesticates. For all its proximity to Papua New Guinea, for example, agriculture failed to reach Australia prior to European contact – or rather, it failed to be adopted there. The fine dividing line between farming and foraging is well illustrated by Australia where aboriginal women customarily exhort plants to be generous and yield a big tuber as they dig them up. Once out of the ground, no matter how large the tuber, tradition decrees that the woman should now complain and berate the plant, ‘Oh you worthless plant, you lazy thing, you stingy plant. Go back and do better’. Saying this, she would chop off

the top of the plant, put it back in the hole from which it came, and urinate on it. This apparently ancient practice resembles deliberate propagation so closely that we are left wondering how the next step – that to true cultivation – could have been avoided (Carter, 1977, p. 95; citing Tindale).

Independent innovation or diffusion?

Much early writing on plant and animal domestication was influenced by diffusionist ideas. Agriculture was thought to be such an extraordinary advance over hunting and gathering that it could only have been ‘invented’ once, from whence it spread to all – or at least most – corners of the Earth (Carter, 1977). As for its place of origin, Griffith Taylor pointed to the Near East, specifically to Egypt, while Carl Sauer believed in a southeast Asian locus. Diffusionist models, in addition to being ethnocentric, also usually invoke intercontinental migration of prehistoric peoples, for instance, from the Old to the New World, or at least close culture contact between them.

One domesticate sometimes used to support the model of cultural dispersal is the bottle gourd (*Lagenaria siceraria*). This holds pride of place among domesticated plants for its early occurrence in areas as distant as Thailand and Mexico. One speculation is that its wide distribution was caused by gourds – which are buoyant – floating from one continent to another. A more likely explanation, however, is that the gourd was a pan-tropical species which, as Jack Harlan (1975, p. 239) put it, is simply ‘the kind of plant that anyone would use’. In particular, it makes an excellent container, especially for water, which would have had obvious advantages for non-sedentary h-f-g populations. Compared with pottery containers, gourds are light to carry, durable and available free of charge from nature. Another pan-tropical domesticate is the coconut (*Cocos nucifera*). This is most likely to have an origin in southeast Asia or the western Pacific, but it was already present on the Pacific coast of the Americas when the Spanish arrived there (Maloney, 1993). Like the gourd, the coconut will float and its fruits can stay viable in saltwater for over three months. Another candidate for early cultural diffusion is the dog (*Canis*), which was present in China, southwest Asia, northwest Europe and North America by the early Holocene (Clutton-Brock, 1977). Significantly, several archaeological sites at which dog is present belonged to mobile hunter-gatherers rather than to agriculturalists. Wild dogs were camp followers of Pleistocene human groups, scavenging for animal bones after kill sites were abandoned (Binford, 1981). Given this reliance on human activities, it is hardly surprising that ‘man’s best friend’ followed the first hunting peoples into the Americas and became our earliest domestic animal (Flannery, 2001). Then, as now, dogs would have acted as camp guards, but in areas such as Mesoamerica, they also provided an item of diet.

Taken overall, bio-archaeological evidence indicates that *global* diffusion of domesticates was rare prior to the last 500 years. Farming systems based on the three great grain crops – wheat, rice and maize – each have independent centres of origin, and the list of plant and animal domesticates for different parts of the world shows little overlap (see [Figure 5.5](#)). American domesticates, for example,

are derived from wild species indigenous to the New World, with species of African and Eurasian origin conspicuously lacking. In fact, it seems remarkable that there are not more than the handful of species, such as the dog and bottle gourd, which are common to more than one region of domestication. This is not to deny that domesticates were exchanged between adjacent regions, such as Peru and Mesoamerica, nor that farming was diffused from southwest Asia into Europe – but it does disprove the idea that agriculture as a global mode of production has a single point of origin. Within any particular centre of origin, genetic evidence suggests that domestication of primary crops like einkorn wheat, chickpeas, rice and maize is most likely to have been accomplished no more than a few times (Zohary, 1989; Blumler, 1992; Heun et al., 1997; Matsuoka et al., 2002; Burger et al., 2008; Molina et al., 2011).

The role of environmental change in early agriculture

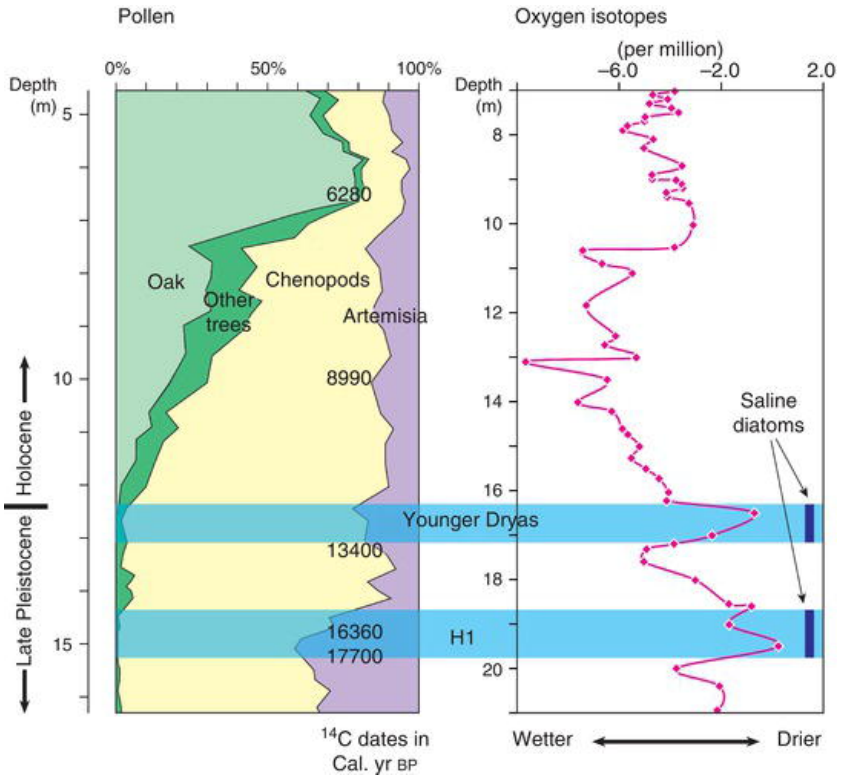
Agriculture is often seen as historically inevitable, but its advantages over a hunter-forager mode of production are not as obvious as they might at first appear. For example, a harvest from dense natural stands of wild einkorn in southeast Turkey yielded a kilogram of clean grain per hour (Harlan, 1967). The grain was of comparable nutritive value to modern cultivated wheat, but involved no expenditure of labour on preparing the ground, sowing or weeding. Consequently, about 50 kcal of energy were netted for every 1 kcal expended, much more than the average of 17 kcal reported for a range of subsistence agricultural systems by Black (1971). There are numerous similar ethnobotanical examples of wild rice and *Panicum* millet gathering (Harlan, 1975, p. 15 ff.; Harris, 1984). We can assume that hunter-gatherers are unlikely to have taken on the additional laborious tasks required for agriculture unless they had a strong incentive to do so. Explanations for the beginnings of agriculture have therefore often involved some shift in the balance between populations and their food resource base, either because of population increase or environmental change.

One of the best-known theories relating climatic change to domestication was proposed by Vere Gordon Childe as long ago as 1929 for southwest Asia. At that time, it was believed that high-latitude glacial periods were accompanied by lower-latitude **pluvials** or wet phases. Evidence of former high water levels in lakes such as the Dead Sea was used to support this glacial-pluvial correlation. Childe logically surmised that the beginning of the Holocene should have witnessed a progressive drying out – or desiccation – of southwest Asia, with water resources increasingly concentrated in locally favoured habitats such as river valleys. In these few ‘oases’, plants, animals and human beings would have been forced to reside in close proximity on a reduced resource base, conditions that rapidly led to new relationships in the form of domestication (Childe, 1935). Childe’s oasis – or propinquity – hypothesis fitted well with archaeological

evidence for an early Holocene ‘Neolithic revolution’ in southwest Asia.

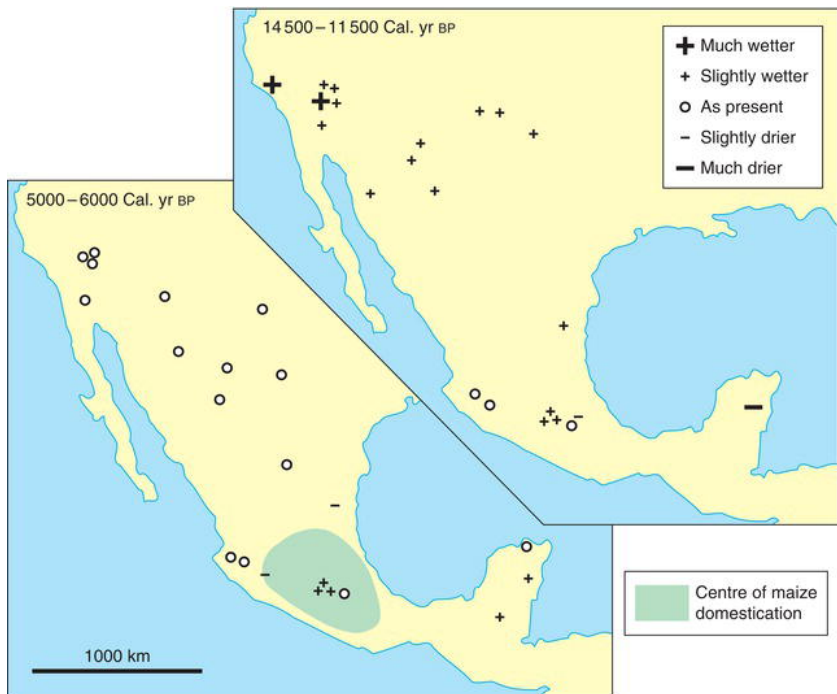
The first serious test of Childe’s hypothesised climatic explanation came with the study of sediment cores from Lake Zeribar in Iranian Kurdistan. Pollen analysis by Willem van Zeist and Sytze Bottema showed that instead of the present oak parkland, the region was treeless chenopod–*Artemisia* steppe before 11.7 ka BP, a flora typical of a cold dry climate, not one of moist climatic conditions. Further pollen studies in southwest Asia have confirmed the Zeribar sequence (van Zeist and Bottema, 1991), and they indicate a sequence of climatic and vegetation changes precisely the reverse of that predicted by Childe’s oasis hypothesis (see [Figure 5.7](#)). During glacial times, wild cereals were common only in a few areas such as the southern Levant, but increases in temperature and rainfall around the start of the Holocene allowed these to expand to their present distribution over the Fertile Crescent (Roberts and Wright, 1993; Wright, 1993). This expansion of the plant and animal resources needed for domestication may have been halted, or even reversed, just prior to the ‘Neolithic revolution’, by the climatic deterioration of the Younger Dryas stadial (13–11.7 ka BP). This climatic reversal led to low water levels and increased salt content in lakes such as Zeribar and Van ([Figure 5.7](#)). A reduced availability of wild cereals and other plant foods at this time might have acted as the stimulus for new forms of human use of these resources in some areas (Moore and Hillman, 1992).

Figure 5.7 Pollen and stable isotope record from Lake Zeribar, western Iran, showing climatic and vegetation changes in southwest Asia during the Late Glacial and early Holocene (data from van Zeist and Bottema, 1991; Stevens et al., 2001). Reproduced with permission from Willem van Zeist.



In Mesoamerica, an oasis hypothesis for plant domestication does not appear particularly consistent with the climatic evidence either. Moisture conditions during the late Pleistocene and Holocene, inferred from pollen diagrams, lake levels, packrat midden macrofossils and other sources, are shown in [Figure 5.8](#). Mesoamerica itself does not show any clear regional climatic trend during the early Holocene (Metcalfe et al., 2000). On the other hand, northwestern Mexico and southwestern United States experienced a very marked desiccation at this time, with water resources concentrated in river valleys, springs and other oases. Yet it was not here but in central Mexico that crop domestication and cultivation began, showing that climatic desiccation on its own was unlikely to prompt a pathway from hunter-foraging to farming.

Figure 5.8 Past moisture conditions in Mexico and southwest United States (adapted from Street-Perrott et al., 1985).



These contrasting explanations illustrate two of the basic models most often invoked to link environmental change to early agriculture. Childe's oasis hypothesis is based on the deterministic idea that necessity is the mother of invention. Deterioration in the natural resource base would lead to food shortage, hence the need for a new mode of production capable of higher economic output. Despite its general unattractiveness to hunter-foragers, agriculture has the capacity to support much higher population densities than economies reliant on wild resources. And while this model does not work particularly well for southwest Asia, China or Mesoamerica, there is a better fit in some other contexts, particularly in Africa where domestication in the Sahelian–Sudanian zone was associated with Saharan desiccation after 5500 years ago (Kuper and Kröpelin, 2006). In any case, determinist models are not always as simple as Childe's, but have also involved population as a dynamic variable (Cohen, 1977; Bocquet-Appel, 2008) or envisaged connections between groups in areas of better and worse resource endowment (Binford, 1968).

In contrast to this model is the opportunist one, in which resources made newly available as a result of environmental change are seized upon by humans who have 'ever had an eye for the main chance'. If rising sea levels put marine food resources on your doorstep, you hunt less and fish more, as the people of Franchthi Cave did at the end of the Pleistocene (see [Figure 4.2](#)). If climatic amelioration creates extensive stands of wild wheat, you learn to harvest them, as the early Natufians did in Palestine around the same time (Rosen and Rivera-Collazo, 2012). Under these circumstances, the question of whether a particular species is domesticated or not can become relatively arbitrary (Donkin, 1997).

How, for instance, should we classify sheep still breeding in the wild, but loosely controlled through the provision of salt? Or wild bulls used to sire domestic female cattle? Human manipulation of plants and animals has an ancestry considerably older than farming proper, and experimentation with plant and animal domestication has to be seen in this wider context (Rindos, 1984; Harris, 1996b). Advanced hunter-foragers – who were no doubt excellent botanists and zoologists – would consequently have probably been surprised at our ‘nitpicking’ distinctions between one form of manipulation and another. Much more important was the change from experimentation with potential domesticates to their widespread adoption in fully agricultural economies. The climatic and environmental changes at the end of the Last Ice Age helped to stimulate a whole range of new ‘experiments’ with plants and animals. However, the *adoption* of farming as a way of life took place sometime after the start of the Holocene, about 2000 years later in southwest Asia, 5000 years in China and 7000 years in Mexico. Climatic change may have played an essential part in allowing farming to emerge during the Holocene, but it was not usually a direct trigger. Other factors, such as feasting rituals, also played a part, as reflected in the fact that specialized religious sites emerged in parallel with permanent settlements. The remarkable ritual complex at Göbekli Tepe in southeast Turkey, for example, was established more than 11 000 years ago, before true farming was in place (Schmidt, 2000).

A problem with both determinist and opportunist models is the large spatial and temporal scales over which they have been applied. Childe’s oasis hypothesis, for example, related to the whole of southwest Asia and covered a time span of many thousand years. For prehistoric communities, however, environmental changes over decades or centuries would have been more important than those occurring over millennia. Similarly, the most important environmental resources were usually those immediately adjacent to their places of habitation, not those averaged over wide areas. The value of studying site-specific resources and environmental changes can be seen from a comparison of where late pre-farming (Natufian) sites and early farming (Neolithic) settlements in southwest Asia were located.

At the western edge of the upper Jordan valley (Figure 5.2), Ain Mallaha is a good example of an open Natufian encampment. It was inhabited around 14–12.5 ka BP at the time of the Bølling–Allerød interstadial and part of the Younger Dryas stadial. Mallaha lies at the junction of several habitats including Lake Huleh surrounded by extensive marshlands and upland hills which supported oak parkland with grasses including wild cereals. It was therefore ideally located to support a broad-spectrum economy. Pollen cores from Lake Huleh indicate that open oak parkland covered the surrounding hills at the time (Baruch and Bottema, 1991), so that plant resources such as acorns and wild cereals would have been available locally to the inhabitants. Phytoliths found in the Natufian layers at Mallaha confirm that wild cereals were being harvested (Rosen, 2004). The presence of shoreline gravels at the archaeological site shows that the community lived near the edge of Lake Huleh, which had expanded to cover what is today valley marshland. Fish, wildfowl, turtles and freshwater shellfish would have then been abundant, and it is not surprising that their exploitation is testified from archaeological excavations, for instance, in the form of fish hooks and net weights.

By contrast, most later pre-pottery and pottery Neolithic sites in southwest Asia lie on, or close to, alluvial soils. They include sites such as Abu Hureyra, on the edge of the Middle Euphrates floodplain, and Çatalhöyük on the flat alluvial Konya plain of the Anatolian plateau (Figure 5.2). These landscapes provided water-retentive soils which helped augment the limited rainfall available for crop cultivation. The success of farming at these ‘mega-villages’ is reflected in the large quantities of charred wheat, barley and other cultigens found during archaeological excavations (Moore et al., 2000; Fairbairn et al., 2002). Geomorphological evidence indicates that deposition of fine-grained river alluvium started only during the early Holocene (Roberts, 1991; Boyer et al., 2006). The appearance of the first agricultural settlements around 10 ka BP therefore has to be set against a background of substantial environmental changes during the immediately preceding millennia.

Sites such as Mallaha and Abu Hureyra had contrasting resources in their site catchments in the past no less than they do today. The switch from a broad-spectrum economy to one based on farming required new resources and hence new site locations. A shift towards alluvial site locations was widespread in southwest Asia and was often coincident with the onset of Holocene alluviation in river valleys. The exploitation of these new alluvial soil–water resources may therefore have represented a form of ‘geological opportunism’ on the part of the first farmers (Vita-Finzi, 1969; Limbrey, 1990).

Early agricultural impacts

Environmental changes associated with the last glacial-to-interglacial transition were undoubtedly instrumental in bringing about the adoption of agricultural modes of production. But the early Holocene brought opportunities to h-f-g peoples; it did not predetermine social and economic evolution. Similar environmental opportunities had existed during previous interglacial periods, but Pleistocene hominins had been unable or unwilling to exploit them. An opportunist model linking environmental change and early agriculture therefore needs to be set in the broader context of post-Pleistocene cultural and ecological adaptations (Binford, 1968). This helps explain why agriculture developed independently in several parts of the world and why it took a different course in each (Barker, 2006). Thus, animal domesticates were an integral part of early agriculture in southwest Asia, but not in Mesoamerica; wild food procurement was abandoned rapidly in southwest Asia, but more slowly in parts of East Asia; nuclear hearth areas existed for mid-latitude cereal domestication, but not for tropical agriculture; and so forth.

In any case, the advent of agriculture was not only influenced by environmental conditions but in turn brought tremendous potential for modifying natural environments. The most immediate impact was on the domesticates themselves. Cultural pressures did not stop at the point of domestication, but continued to select the most productive forms, which in crops has produced a general increase in size, especially of the plant organ used by people. Many

modern cultigens, such as maize, are consequently hardly recognisable as the descendants of their wild ancestors (see [Plate 5.3](#)). Domesticates often also lose their natural means of dissemination and become dependent on agricultural practices for their propagation. But if cultigens could not survive without human help, neither could humankind survive without cultivated crops. They have domesticated us as much as we have domesticated them.

Agriculture thus strengthened the mutual dependence between people (as farmers) and a limited range of (domestic) plants and animals. In so doing, it also brought humans into conflict with other elements of nature. The artificial ecosystems created by agriculture protected crops and farmyard animals from predators and advantaged them in relation to natural competitors. The consequences were sometimes greater for these wild species than they were for the domesticates themselves. Sheep and cattle are numerous in modern Britain, after all, while predators such as the wolf (*Canis lupus*) and competitors such as the aurochs (*B. primigenius*) are now extinct, either locally or globally. Extinction has occurred both directly through hunting kill-off and indirectly through loss of habitat. The latter relates to the most important aspect of all in the new human relationship with nature – farming's use of the land. Cultivation of crops requires at least partial clearance of the existing vegetation cover by tree felling, ring-barking or burning and its replacement by field plots. Consequently, early agriculture is associated with the first substantial human impact upon the soil. This impact was all the more permanent because of agriculture's association with a settled, or sedentary, way of life.

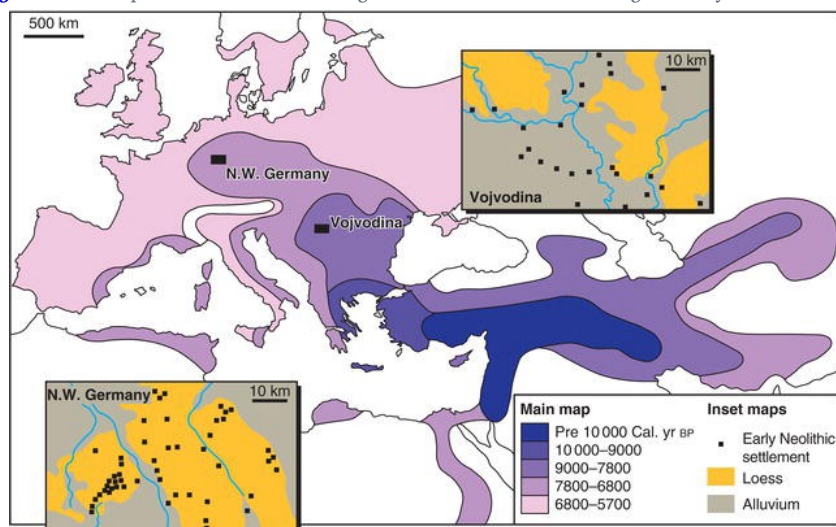
Sedentism is known in a few non-farming societies such as the Kwakiutl culture of Canada's Pacific coast. Fully sedentary communities probably existed around the former lake shores of the basin of Mexico from 9 ka BP on, over four millennia before agricultural villages developed there (Niederberger, 1979). However, in both cases, sedentism was only possible because of locally favourable habitats in which resources such as wildfowl and fish were available year-round. It is similarly true that some agricultural societies, notably those employing shifting cultivation, can exhibit varying degrees of mobility. Nonetheless, it remains the case that a sedentary way of life is much easier under a food-producing economy than it is for one based on foraging or hunting. Cultivated fields require attention through much of the agricultural year, while the need to store harvested crops also reduces freedom of movement. Peasant farmers are, in a real sense, tied to the land. Some of the best recorded examples of early agricultural impacts come from Neolithic Europe, and these are used here to illustrate the diffusion and environmental relations of the first farming communities.

European agricultural dispersals

Between 9 and 5.5 ka BP, Neolithic farming spread across Europe from southwest Asia, westwards via the Mediterranean and northwestwards along the Danube–Rhine axis (see [Figure 5.9](#)). Most of the latter was accomplished in two great leaps in the frontier of settlement: one from the Aegean to the Great Hungarian

Plain (8.2–7.8 ka BP) and a second from there across the North European Plain (c.7.4 ka BP). The sharp discontinuity between late Mesolithic and primary Neolithic site locations and economies in central Europe suggests that here the initial spread involved colonisation by pioneer Neolithic farmers, so-called demic diffusion (van Andel and Runnels, 1995), rather than by acculturation. This is a conclusion supported by the study of human DNA, both ancient (e.g. Bramanti et al., 2009) and modern (e.g. Balaesque et al., 2010), at least for the male line. At sites such as Franchthi Cave, where there was continuity of occupation through the Mesolithic–Neolithic transition, the food procurement system changed abruptly at the Mesolithic–Neolithic transition (Hansen, 1991), suggesting that new people had arrived. Some authors have seen an analogy with the frontier of European settlement in North America during the eighteenth- and nineteenth-centuries AD and have tried to measure the rate at which the population ‘wavefront’ advanced across Europe (Ammerman and Cavalli-Sforza, 1971; Bocquet-Appel et al., 2009). However, Neolithic farmers only exploited a small portion of the total landscape, selecting those particular habitats – notably alluvial and loess soils – best suited to their agroecosystems (Figure 5.9). In the western Mediterranean and parts of northern Europe (e.g. the Ertebølle culture in Denmark), sites with transitional economies do occur, indicating that here agriculture is more likely to have been adopted by pre-existing Mesolithic populations. Environmental and cultural factors, as well as demographic ones, therefore determined the rate at which early farming spread (Dennell, 1983; Bugucki, 1996; Zvelebil and Lillie, 2000; Bonsall et al., 2002; Lemmen et al., 2011). The end result was a dramatic increase in Europe’s human population between Mesolithic and Neolithic times, probably by an order of magnitude to judge from the number of ^{14}C dates available for each archaeological period (Shennan and Edinborough, 2007).

Figure 5.9 The spread of Neolithic farming across western Eurasia during the early Holocene.



The first farmers of southeast Europe operated agroecosystems strongly reminiscent of those developed in southwest Asia. Sheep/goat and emmer wheat were economic staples, but now they were living outside their natural ecological zones and could only survive with human protection. The crops on which the system depended were cultivated by intensive garden horticulture on hydromorphic alluvial soils (Sherratt, 1980). These alluvial environments were attractive to early agriculturalists because they provided a reliable supplementary source of soil moisture in a climate where rainfall was, as it still is, one of the main factors limiting plant growth. Cereals, once transferred from their natural habitats on to water-retentive soils, raised the carrying capacity of one small section of the environment out of all recognition. As in southwest Asia, most alluvial environments in southeast Europe had only taken on their modern form early in the Holocene (van Andel and Runnels, 1995).

The natural environment placed limits on the form that Neolithic agroecology took, and once it reached the Great Hungarian Plain, it encountered environmental conditions sufficiently different from those in southwest Asia that it was forced to re-orientate. The contrast in Neolithic settlement locations between one side of the Hungarian Plain and the other is very marked, with alluvial soils used on the east and loess ones on the west (Kosse, 1979). The Neolithic Linear Pottery (LBK) culture found in western Hungary subsequently spread with great rapidity through central and northern Europe (Keeley, 1992; Whittle, 1996). LBK cultivation remained based on cereals and pulses (Bogaard, 2004; McClatchie et al., 2013), but animal husbandry switched from sheep and goat to a reliance on cattle and pig – whose wild progenitors could be found in temperate Europe (see [Plate 5.4](#)). Alluvial soils were no longer required because moisture deficiency was not a clear limiting factor for crop growth in central and northern Europe. But why soils on [loess](#) were so deliberately selected as their replacement is less clear. Loess can certainly produce fertile chernozem soils with a large nutrient store and a loamy texture suitable for hoe cultivation. These soils may also have been chosen because they were only rarely occupied by pre-existing Mesolithic groups, as shown by detailed archaeological surveys, such as that on Germany's Aldenhovener Platte (see [Figure 5.9](#), inset map). Expanding Neolithic farmers may therefore have been sucked into the empty portions of the landscape.

Plate 5.4 Wild boar (*Sus scrofa*) from central Europe: the ancestor of the modern farmyard pig.



There has been debate about whether the first farmers of temperate Europe operated an extensive system of land use based on shifting – or swidden – cultivation or continued to rely on a more intensive garden horticulture (Bogaard and Jones, 2007; Kreuz, 2008). A shifting agro-ecological system has been practised in some of Europe's boreal forests until recently, for example, in northern Finland (Lehtonen and Huttunen, 1997), although its most widespread occurrence is in the humid tropics. Evidence for shifting cultivation by European Neolithic farmers comes from their great wooden long houses, each occupied by a single extended family and their livestock. Unlike the long-lived and permanently occupied village tells of the Balkans and southwest Asia, the central European long houses were only occupied for a limited number of years before being abandoned. At the LBK site of Bylany in Bohemia, a 50-year agro-ecological cycle has been proposed, with each house site occupied for 10–15 years each (Soudsky and Pavlů, 1972). This cycle would have been repeated many times in any one area, with initial forest clearance involving burning as well as tree felling, partly to increase soil nutrients. After cultivation ceased, shrub and some tree vegetation recolonised the land allowing soil fertility to recover.

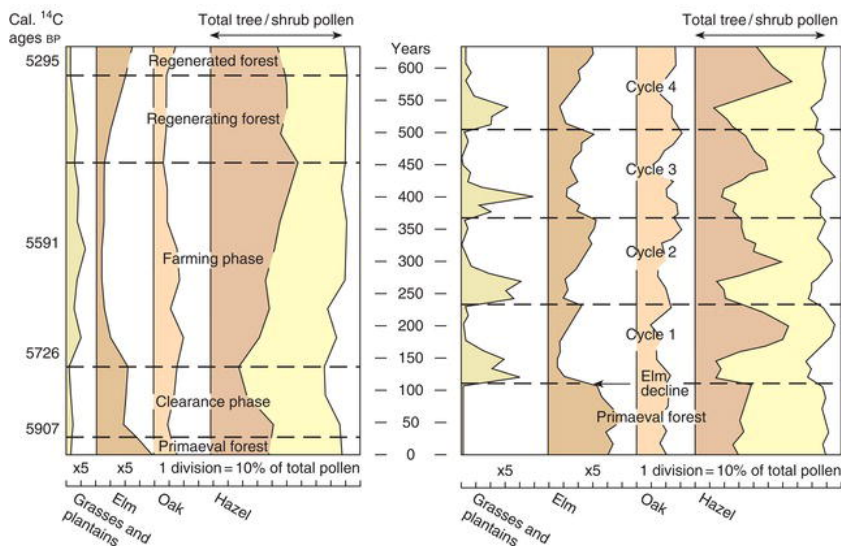
Ecological consequences of early European agriculture

A swidden agro-ecological system might be expected to have had a more intensive impact on native European forests than Mesolithic woodland management and a wider one than the alluvial garden horticulture of southeast Europe, neither of which has left a clear mark in most pollen diagrams (Willis and Bennett, 1994). So to what extent are vegetation clearance, cereal cultivation and woodland

regeneration indicated in mid-Holocene pollen diagrams, and do they support the 50-year long fallow cycle suggested at Blyany? In a paper published in 1941, Johannes Iversen was the first to recognise clearance or, as he called them, *landnám* phases in pollen diagrams. In some respects, his *landnám* pollen phases conform well to the 'slash and burn' shifting cultivation model. They include an initial clearance stage in which tree pollen declines relative to herb and grass pollen, a farming stage in which grasses including cereal-type and weedy species reach a maximum and a regeneration stage in which shrubs such as hazel increase before declining as more substantial trees replace them. The clearance phase is also sometimes associated with a rise in the frequency of charcoal, suggesting that fire was employed in a 'slash and burn' manner.

In other respects, however, pollen and archaeological data have produced some surprising results. Iversen originally estimated a duration of 50–100 years for the *landnám* cycle, but subsequent ^{14}C measurements have shown that the first Neolithic demographic cycle lasted much longer, generally between 200 and 600 years. Furthermore, the clearance cycle is rarely repeated in Neolithic pollen diagrams (Shennan and Edinborough, 2007). Swidden agriculture may in fact have involved a short 5–10-year cycle, rather than a long 50-year one, but this is hard to identify in most pollen diagrams where the sampling interval between pollen spectra is usually more than 10 years. Short fallow cycles would simply become blurred together into a general phase of vegetation disturbance (Edwards, 1979, 1993). High-resolution pollen diagrams are not always located close to Neolithic settlement sites, and they are virtually absent from drier soil types such as loess. The contrast between an actual *landnám* phase, as recorded in a pollen diagram from Ballyscullion in Ireland, and that expected under long fallow shifting cultivation is shown in [Figure 5.10](#).

Figure 5.10 Pollen diagrams recording: (left) actual *landnám* phase at Ballyscullion, Ireland (data from Pilcher et al., 1971; Smith et al., 1971); (right) sequence expected under long fallow shifting agriculture.



The effect of Neolithic clearance on the overall woodland cover – although clearly detectable (Woodbridge et al., 2013) – was initially rather small, with tree pollen continuing to be dominant right through *landnám* events. Indeed, Ian Simmons and John Innes (1996) suggest that the first cereal growing may have been slipped into patches of land which had been cleared of other purposes, such as attracting deer. More significant were changes that took place in the composition of both woodland and open-habitat vegetation (Whitehouse, 2006). One of the species affected was the understorey shrub hazel (*Corylus*). Hoe-based shifting cultivation does not require felling of large trees, which can be left standing inside field plots. By contrast, hazel scrub was cleared during cultivation phases, but would have regenerated quickly in any succeeding – and longer-lasting – fallow periods, and it thrived in the open, well-lit conditions. Hazel can survive the repeated firing of ‘slash and burn’ because it coppices freely, and the hazel rods produced from the shoots were of considerable practical value. They were, for example, the main source of wood used in the buried Neolithic trackways of the Somerset Levels of southwest England (Coles and Coles, 1986). Higher, but rather fluctuating, levels of hazel pollen characterise many northwest European woodlands in early and mid-Holocene times (Huntley, 1993), and it has been suggested that this was a result of prehistoric cultivation and woodland management (Smith, 1970). Hazelnuts are found commonly in archaeological sites from Britain, not only during the Mesolithic but also during Neolithic times (Stevens and Fuller, 2012), implying that they were an important food source even after the introduction of cereal agriculture. Hazel was therefore almost certainly being exploited as a managed renewable resource.

One species whose importance in European pollen diagrams changed even more markedly than that of hazel during Neolithic times is the elm tree (*Ulmus*). The sharp and often permanent decline in elm pollen around 5.8 ka BP has probably received more attention than any other feature of European vegetation

history. Numerous hypotheses have been proposed to explain this arboreal enigma, although they can be grouped together under four main headings (Peglar and Birks, 1993; Parker et al., 2002):

- 1 Catastrophic pathogen (disease) outbreak;
- 2 Anthropogenic disturbance;
- 3 Competitive exclusion by later migrating tree species;
- 4 Unfavourable climatic changes.

Although many – or all – of these could have operated in tandem, the first two hypotheses are most often invoked. The possibility of a pathogen outbreak has obviously been placed in the forefront of botanical consciousness as a result of Dutch elm disease which decimated Europe's elm population during the late twentieth century (Perry and Moore, 1987). The elm bark beetle (*Scolytus*), which is the main carrier of the fungus responsible for Dutch elm disease, was certainly present in Britain during the mid-Holocene (Girling and Greig, 1985). Furthermore, in fine-resolution pollen study of laminated sediments from Diss Mere in eastern England, Sylvia Peglar (1993) showed the elm decline to have been extremely rapid – only six years at this site – which is consistent with a catastrophic pathogen attack. On the other hand, there is a strong chronological coincidence between the elm decline and early Neolithic farming, at least in northwest Europe. As at Ballyscullion, the elm decline typically accompanies the beginning of *landnám* phases and is associated with pollen, such as cereal type, indicative of human disturbance. Anthropogenic explanations do not generally envisage large-scale felling of elm woods to make way for Neolithic agriculture. Instead J. Troels-Smith (1960) proposed that elm branches were lopped to provide winter fodder for stalled cattle, so reducing production of elm pollen, but not necessarily greatly reducing the number of elm trees. Pollarded trees would also have been more susceptible to attack by the disease and its carriers. So while disease is most the probable immediate cause of the tree's decline, it is likely that elm populations would have eventually recovered, if it had not been for the additional impact of Neolithic farmers.

A third group of plant species to respond to Neolithic agriculture were weeds (or ruderals). Species such as ribwort plantain (*Plantago lanceolata*), stinging nettle (*Urtica dioica*), docks and sorrels (*Rumex* spp.) and many grasses (Gramineae/Poaceae) regularly appear in post-Mesolithic pollen diagrams. These plants thrive on disturbed ground, so that they were pre-adapted to survive in culturally disturbed habitats, and the spread of agriculture was paralleled by the spread of weedy species (Behre, 1986). These opportunistic plants exploited human agency for their dispersal and have remained a familiar – if annoying – part of fields and gardens ever since (Mabey, 2010).

One of the ironies of the European pollen record is therefore that the best indicators of early agricultural disturbance are not cultivated crop plants, but the elm tree and ribwort plantain! Both testify how the ecological consequences of Neolithic farming were largely indirect, even unintended. The vast majority of Europe's wildwoods remained untouched by the initial arrival of agriculture, and those that were exploited formed part of an integrated agro-ecological system. The first farmers needed woodland, not only for timber and fodder but also to maintain the productivity of their cultivated crops. They depended on a fertile

soil and clement weather for their crops and good graze and browse for their livestock. Degradation of those environmental resources would have undermined the system of food production and threatened their own survival. Consequently, while the development of farming may have changed the relationship between people and nature, the fate of agricultural societies remained interwoven with the habitats they occupied.

Conclusion

Farming began independently on at least four separate occasions during the first half of the Holocene in centres as far apart as Mesoamerica and Papua New Guinea, having failed to do so anywhere during the preceding Pleistocene period. This suggests that farming was only likely to be developed by anatomically modern humans with advanced h-f-g economies living in an interglacial world. Such a combination occurred for the first time around 11 000 years ago. Prior to that time, the glacial climate was not only cold and hostile but also frequently unstable and therefore unreliable (see [Figure 3.3](#)). In turn, this would have discouraged the intensified use of wild plant resources prior to the more stable climate of the early Holocene (Richerson et al., 2001). When farming did develop, it did so because of locally suitable circumstances and, most importantly, because of the occurrence of large-seeded wild cereal grasses such as einkorn wheat (Blumler, 1996). For communities living nearby, there were good reasons to invest time and energy in highly edible and easily stored seeds to create monospecific stands which in time became fields of cultivated crops. These human groups were given a head start over the rest of the world, a lead which has been largely maintained ever since (Diamond, 1997). During the second half of the Holocene, Mesoamerica, southwest Asia and East Asia all gave birth to complex cultures and civilizations, such as the Aztecs, Sumerians and Han. At the root of this success is good and abundant food, and it is reflected in Mexican, Mediterranean and Chinese cooking still being among the best in the world. The way that the die was cast early in the Holocene had human consequences which remain of great importance today.

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The taming of nature (6000–1000 Cal. yr bp)

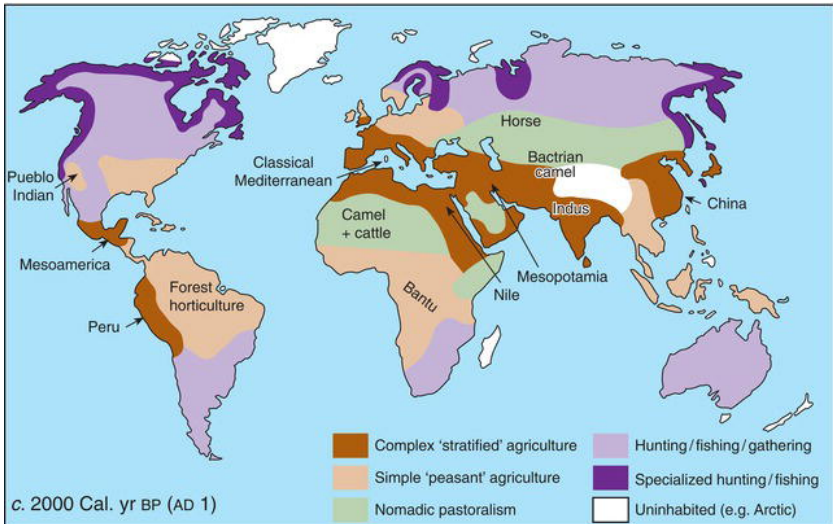
Introduction

The middle and later part of the Holocene was the period in which the balance of power for the control of nature permanently shifted. The world some 5000 or 6000 years ago was still primarily a product of natural forces – climate, relief and so forth. It was much as it appears in the front of atlases with natural regions of forest and grassland, ice and ocean. Where it differed from today's 'natural regions' – for example, the absence of today's arid Sahara – this was because of subsequent natural climatic changes, not because of human impacts. During the early Holocene, humans can reasonably be considered to have formed a part of the natural world or, to take an alternative view, were dominated by nature (Simmons, 1996). This relationship derives from the fact that the majority of the world's human population were still reliant on hunting, fishing and gathering (h-f-g), a mode of production which normally uses and manipulates natural ecosystems but does not transform them into something else. It is true that peasant farming societies had begun to emerge prior to 6000 years ago (see Chapter 5), but these were as yet restricted to small parts of Europe, North Africa and Asia. The long process of clearing natural vegetation to make way for cultural landscapes was only just beginning.

Contrast this situation with that at the end of the time period covered in this chapter. Thousand years ago, many natural ecosystems had been replaced by agricultural ones (Kaplan et al., 2009, 2011; Ellis et al., 2013) and even those retaining the vestiges of their original appearance had been altered in their detailed composition. Out of primary nature, human agency had fashioned what Clarence Glacken (1967) termed a 'second nature'. This taming of nature was the work of agricultural societies which developed and spread over large parts of all the major land masses except Australia and Antarctica (see [Figure 6.1](#)). The spread and development of agricultural modes of production created cultural landscapes such as the olive groves of the Mediterranean and the irrigated rice fields of China's river valleys. Cultural changes did not have a uniform effect over the globe, however, because the pace and direction of cultural evolution varied

from one region to another (Simmons, 1996; Diamond, 1997). Furthermore, the natural environment responded dynamically to human influence, sometimes metamorphosing in a way quite different from that intended by the initial human actions. On the Pleistocene sands of northern Europe, for example, forest clearance led to soil podzolisation and the creation of heathlands rather than reversion to secondary woodland (Gimingham and de Smidt, 1983). It is consequently at the regional scale that human–environment relations are most usefully examined during most of the mid- to late Holocene (Dearing et al., 2006). For this reason, much of this chapter is devoted to selected regional environmental histories: for example, in Mesoamerica, the Mediterranean basin and the British Isles.

Figure 6.1 Global cultural developments during the later Holocene (adapted from Lewthwaite and Sherratt, 1980).



Landscapes were not transformed by human agency alone, however, because nature was actively changing independent of human action. Natural agencies such as fire, disease, sea-level change and soil maturation all continued to operate, and the climate was far from stable. Environmental change associated with this array of natural factors is the first theme to be explored in this chapter.

Changes in the natural environment

Climate and vegetation

Although much less marked than during the early Holocene, the second half of the Holocene nonetheless experienced some significant shifts in climate (Wanner et al., 2008; Marcott et al., 2013). Among these was a progressive cooling in many regions following a Holocene thermal optimum around 9–5.5 ka BP. This deterioration was particularly marked in central and northern Europe (Leemann and Niessen, 1994; Heikkilä and Seppä, 2003) and adjacent areas of the North Atlantic (Anderson et al., 2004; Caseldine et al., 2006) and often involved a series of steps rather than a gradual decline. One of these steps was from the warmer and drier Sub-Boreal to the cooler and wetter Sub-Atlantic period, a change dated in European peat bogs to around 2.6 ka BP (see Table 4.2). The effect of late Holocene cooling on vegetation belts in Europe is often disguised by increasing human disturbance. The northward spread of beech (*Fagus*), for example, may have been partly due to climatic cooling which disadvantaged competing thermophilous species such as the lime tree (*Tilia cordata*). But beech was also aided by Bronze and Iron Age clearance, which provided gaps in the forest and allowed it to invade and become a dominant tree over large parts of northwest Europe (Huntley and Birks, 1983, p. 191ff.; Huntley et al., 1989; Küster, 1997). In North America, where the pace of cultural evolution was slower, there is less ambiguity about the relative effects of natural and human agencies. The westward shift in the prairie–forest boundary and the southward movement of the boreal forest after c.4.5 ka BP both reflect the response of vegetation to a cooler, moister climate (Webb et al., 1993).

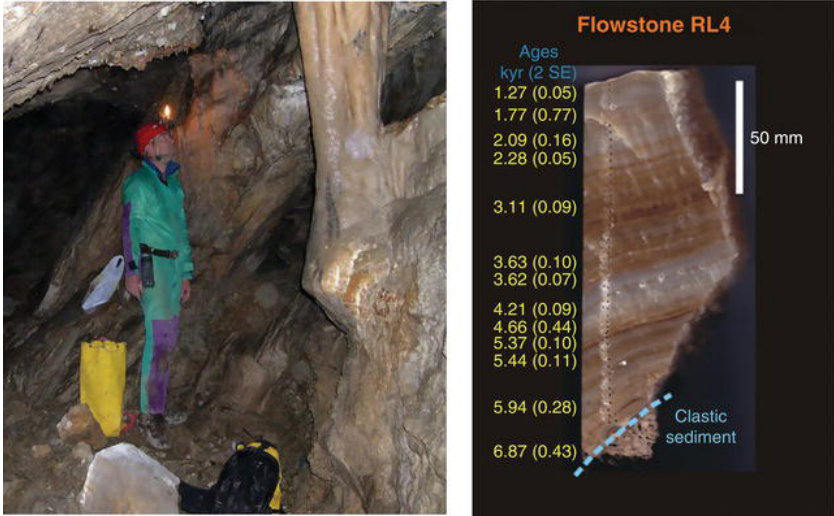
An even more important climatic change occurred in the sub-tropics of the Old World, where climatic desiccation between 6.5 and 4.5 ka BP created the modern Saharan, Arabian and Thar deserts (Hoelzmann et al., 2004). Pollen records from the Sahara (e.g. Ritchie et al., 1985; Kröpelin et al., 2008) show a relatively gradual transition from savanna to semi-arid scrub to barren desert over this time period. In contrast, the water levels of some low-latitude lakes fell abruptly, suggesting that the monsoonal circulation system which controls tropical rainfall was suppressed much more rapidly (Street-Perrott et al., 1991; Eggermont et al., 2008). Peter deMenocal and colleagues (2000) found an abrupt increase in windblown dust in deep-sea cores off West Africa around 5.5 ka BP, suggesting that desertification occurred rapidly in the Sahara, possibly involving a positive feedback loop. If correct, then this may represent a tipping point in the Earth system in which vegetation cover and hydrological recycling via wetlands across North Africa no longer contributed to regional rainfall. Not surprisingly, the change from savanna to desert completely altered the human ecology of the Saharo-Arabian zone. Without lakes or rivers, the ‘Aqualithic’ culture of the early Holocene disappeared (see Chapter 4), while in the few remaining oases, such as the Nile valley, new intensified forms of land use emerged (Hassan, 1986; Kuper and Kröpelin, 2006). The slowing of eustatic sea-level rise had already reduced the Nile river’s gradient and led to silt accumulation and the creation of the fertile delta plain (Stanley and Warne, 1993). As the level of the Nile flood declined between 6 and 4 ka BP (Butzer, 1984), simple floodwater farming became artificially regulated. New water lifting technology such as the *shaduf* was progressively introduced, and this had major consequences for social and political

conditions in Egypt (Butzer, 1976; Hassan, 1993; see following text).

The middle part of the Holocene also saw the establishment of the modern ENSO regime around the Pacific (Markgraf and Diaz, 2000; Cane, 2005; Chiang, 2009). El Niño events are known to have occurred even in glacial times, but spectral analysis of varved lake sediments from Ecuador shows that their frequency and amplitude was significantly different during the early Holocene compared to recent millennia (Rodbell et al., 1999; Moy et al., 2002). Other proxy-ENSO climate records come from Pacific corals (Gagan et al., 2004), from mollusc species found at Peruvian archaeological sites (Sandweiss et al., 2001) and from pollen records in eastern Australia (Donders et al., 2005). Together they indicate that key changes occurred at around 5 ka and again at 3 ka BP towards more active ENSO cycles in the equatorial Pacific, comparable to the high-amplitude fluctuations in weather that have occurred in recent decades. This shift may have been linked to a gradual change in the temperature gradient between the North and South Poles and the equator, which led to a southward migration of the ITCZ through the Holocene (Haug et al., 2001).

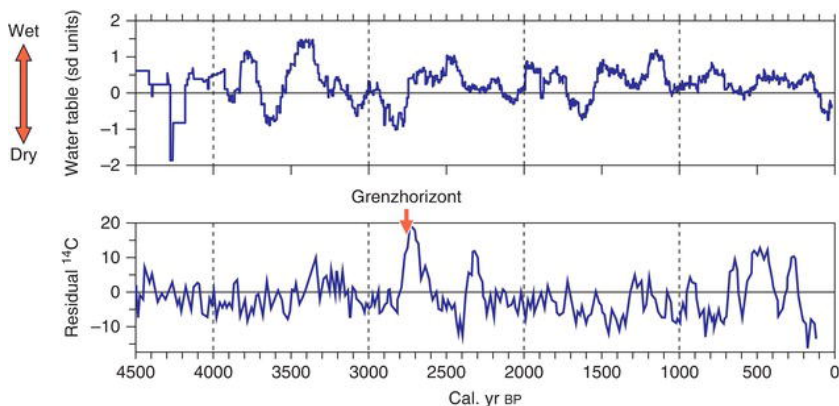
Superimposed on the longer climatic trends related to Milankovitch astronomical changes were shorter-term, sub-millennial shifts from wet to dry or cold to warm (Mayewski et al., 2004). Some of these perturbations to the climate appear similar to, although smaller in amplitude than, earlier abrupt events such as that which occurred 8200 years ago. At around 4.2 ka BP, for example, there was catastrophic failure of the Nile floods which helped to dislocate Old Kingdom of Egypt economically and politically (Stanley et al., 2003). In the inscription of Ankhtifi from the late third millennium BC, it was written that ‘All of Upper Egypt was dying of hunger, to such a degree that everyone had come eating his own children...’ (Bell, 1971, p. 9). There is evidence that winter Mediterranean rains declined at this time (Bar-Matthews et al., 1997; Cullen et al., 2000; Drysdale et al., 2006) (Plate. 6.1), coinciding with the abandonment of some urban centres of the Akkadian Empire (Weiss et al., 1993). However, in other parts of southwest Asia, the same ‘mega-drought’ prompted not cultural collapse but new innovation, particularly in the form of technology for water storage and irrigation (Rosen, 2007).

Plate 6.1 Renella Cave, Italy with U-Th dated speleothem. The white coloured calcite layer dating to 4.2 ka BP marks a drought phase which affected much of the Old World (Drysdale et al., 2006). Reproduced with permission of the Geological Society of America.



Some of the decadal- to century-scale climatic changes of the last 5000 years appear to have been cyclical in character. In an analysis of recurrence surfaces in Danish raised bogs, Bent Aaby (1976) found that oscillations between cool-wet and warm-dry conditions took place every 260 or more infrequently every 520 years since the mid-Holocene. Dan Charman (2010) identified a similar c.280-year periodicity in water-table fluctuations from peat bogs in northern Britain during the last 4000 years (Figure 6.2). Century-scale periodicities have also been recorded from lake-level fluctuations in Europe and Mesoamerica (Hodell et al., 2001; Magny, 2004) and from Holocene histories of glacier advance and retreat (Grove, 2004). One of these glacier fluctuations helped to preserve the body and belongings of Ötzi, the Neolithic ice-man (Technical Box VII). Gerard Bond and colleagues (1997) identified sand layers in North Atlantic deep-sea cores that had been carried in icebergs from Greenland; these IRD (ice-rafted debris) layers reach a peak approximately every 1500 years during post-glacial times. It is uncertain whether the IRD peaks represent true, regular cycles, as is the question of the underlying causes of these climatic changes. Among the hypotheses proposed for decadal- to century-scale climatic fluctuations are explosive volcanic eruptions with large emissions of dust and gases, solar cycles and internally generated stochastic variations in the ocean-atmosphere system (Nesje and Johannessen, 1992; Bradley, 2003; Beer and van Geel, 2008; Wanner et al., 2008).

Figure 6.2 Peat surface wetness record for upland Britain (data from Charman, 2010) and residual ^{14}C as an indicator of solar variability (data from INTCAL04; Reimer et al., 2004).



Of these mechanisms, solar cycles are the most likely to have generated a regular periodicity, for example, related to sunspot activity. The direct radiative effect of solar changes is tiny – less than 0.1% difference between peaks and troughs in the 11-year sunspot cycle – but mechanisms have been proposed by which this signal may have been amplified within the Earth system (e.g. Shindell et al., 2001). Documentary evidence for sunspot activity extends back about four centuries, so that proxy evidence for solar variability is required for older time periods, one of the most useful being short-term fluctuations in the ^{14}C content of the atmosphere. These fluctuations first came to light as part of the cross-comparison between tree-ring ages and ^{14}C dates (see Chapter 2). The residual values – that is, the departures from the long-term trend – are likely to have been caused by changes in the rate of ^{14}C production in the upper atmosphere associated with higher or low solar irradiance, lower $\Delta^{14}\text{C}$ indicating more incoming solar radiation. Trends and cycles evident in the record $\Delta^{14}\text{C}$ residuals (e.g. 206-year-long cycle) can then be compared against similar trends and cycles in proxy climate data. Sometimes these seem to match up quite well. For example, the 1500-year North Atlantic Bond ‘cycles’ have been matched against solar variability (Bond et al., 2001), while the shift in northern European peat bog stratigraphy around 2600 Cal. yr BP occurred soon after a sharp rise in $\Delta^{14}\text{C}$ (Van Geel et al., 1996; Mauquoy et al., 2008; Plunkett and Swindles, 2008; see [Figure 6.2](#)). On the other hand, the matchup is sometimes less obvious. Critical to any test is the reliability of the chronologies being compared, and stretching (or tuning) one to fit the other can easily lead to circular reasoning.

TECHNICAL BOX

Technical Box VII: Ötzi – the ice man

Found by mountaineers in 1991 melting out of the Schnalstal glacier on the border between Austria and Italy, Ötzi is Europe’s oldest natural human mummy, dated by ^{14}C to 5.2 ka BP (Spindler, 1994). It is a very

rare case in which mummification took place by dehydration before the body became embedded in glacier ice. Equally unusually, his body was not carried down valley and deformed by the glacier but instead preserved in the ice at the place where he died (Plate 6.2). In addition to his extensively tattooed body, Ötzi's clothes and possessions were also preserved *in situ*, and they give us an insight into many aspects of daily life in Neolithic times, including people's use of natural resources. For example, he wore a cloak made of woven grass, a bearskin cap, and a vest, a belt, leggings, loincloth and shoes all made of leather (Bortenschlager and Oeggel, 2000; Acs et al., 2005). Other possessions included a copper axe with a yew handle, a flint knife with an ash handle, along with a quiver of 14 bone-tipped arrows with viburnum and dogwood shafts and flint heads. Analysis of Ötzi's intestinal contents showed that his last meals included chamois and red deer meat, processed einkorn wheat bran (possibly in the form of bread) and sloes, a small plum-like fruit of the blackthorn tree. Equally remarkably, pollen analysis from his gut has been used to demonstrate his final journeys (Oeggel et al., 2007). During the last 33 hours of his life Ötzi walked from high up near the timber line (at about 2500 m), to low down in the zone of warmth-loving trees (about 1200 m or less), and finally very high in the zone of perennial ice above 3000 m. Perhaps inevitably, Ötzi has become a subject of controversy, including the question of which country should 'own' him; (the answer turned out to be Italy, since he lay 90 m inside their frontier...and he now resides in the South Tyrol Museum of Archaeology in Bolzano). Equally controversial was the cause of Ötzi's death, an argument now effectively resolved by the discovery via forensic radiological investigation of an

arrowhead inside an unhealed wound under his left shoulder (Gostner and Egarter-Vigl, 2002). In short, Ötzi was shot and eventually killed by an arrow.

Plate 6.2 Ötzi, the Neolithic ice man; (left) the body as found protruding from an Alpine glacier in 1991 (reproduced with permission of the Austrian Police); (right) reconstruction (South Tyrol Museum of Archaeology – www.iceman.it).



Well-preserved ancient human bodies have been discovered in a wide variety of other circumstances around the planet. Mummified bodies have been found in desert environments, such as the Atacama desert of Peru and northern Chile, and in central Asia, the latter including the ‘Siberian Ice Maiden’. Dating to around 2400 years ago, she was found with along with six decorated horses, and her body covered with vivid blue tattoos of mythical animal figures. Very different conditions have led to the preservation of bog bodies found in peat deposits from Britain, Denmark and elsewhere in northern Europe; they include Tollund Man and Lindow Man, both of whom had been ritually executed.

Volcanic aerosols provide an alternative mechanism for disturbing the Earth’s climate, notably by creating a dust veil which reduces the transparency of the atmosphere. After major historically recorded eruptions, such as those of Tambora in Indonesia (1815) and Laki in Iceland (1783), contemporary writers recorded severe late frosts and peculiar hazes in regions far removed from the volcanoes themselves (Grattan and Brayshay, 1995) ([Plate 6.3](#)). Benjamin Franklin, who was

American ambassador in Paris during the 1780s, described a ‘constant dry fog on which the rays of the sun seemed to have little effect’. Some earlier Holocene eruptions were larger than Laki and Tambora (Table 6.1), and it has been suggested that they may have had a proportionately bigger climatic impact. For example, in about 1628 BC, the Aegean island of Santorini (Thera) exploded catastrophically, forcing its Bronze Age population to flee. Dendroclimatological studies have shown that the tree rings of Irish bog oaks were unusually narrow at this time (Baillie and Munro, 1988), while bristlecone pine growth rings of the same age in California show signs of frost damage (LaMarche and Hirschboek, 1984). These have been interpreted as the result of a cooler and/or wetter climate. However, in Turkey, which is normally warm and dry in summer, the same climatic change gave rise to more, not less, favourable growing conditions and to broader tree rings (Kuniholm et al., 1996). A similar cluster of narrow tree rings indicative of adverse climatic conditions occurred around AD 536 – the year that Merlin died according to Arthurian legends. There is debate about the whether the climatic cause of this ‘year without summer’ was volcanic (Larsen et al., 2008) or linked to a comet hitting Earth (Baillie, 1999). The evidence linking tree rings, climate and volcanic eruptions is impressive, but it also suggests that in most cases the climate perturbation was short lived, typically a decade or less (Sadler and Grattan, 1999; Robock, 2000).

Plate 6.3 The 1970 eruption of Hekla in Iceland. Previous eruptions left tephra marker layers across northern Europe and had major ecological consequences on Iceland itself. Photo credit: Oxford Scientific Films (Photo: Hjalmar R. Bardarson).



Table 6.1 Some major volcanic eruptions during the last 7000 years, identified in the GISP2 ice core record

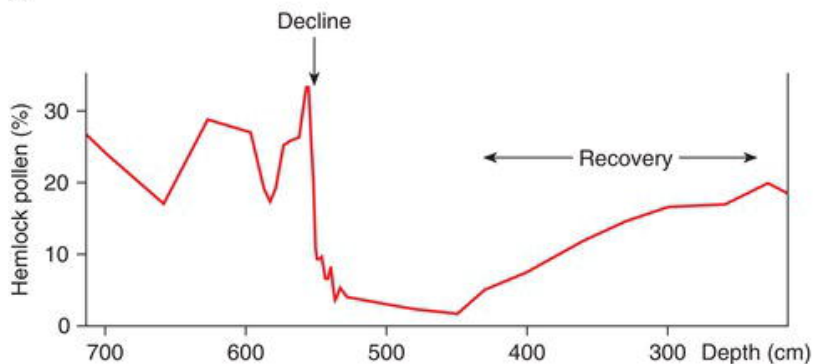
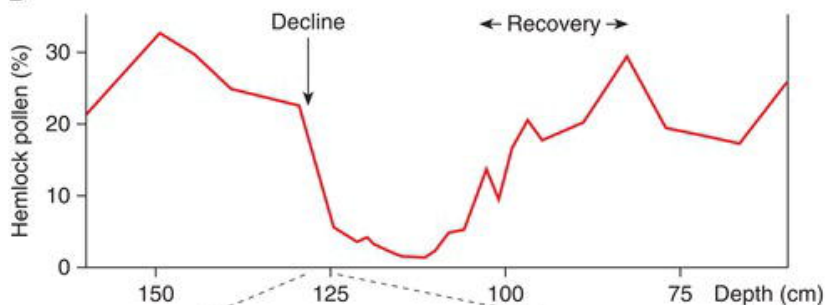
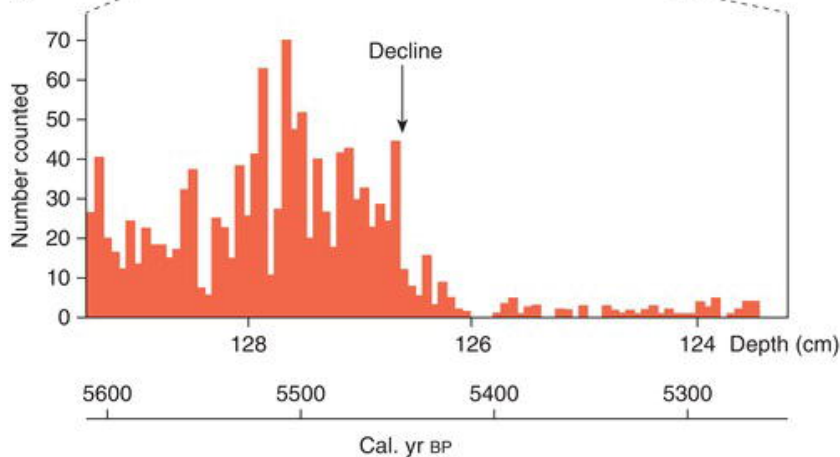
Date	Volcanic eruption	SO ₄ anomaly (ppb)	VEI
AD 1815	Tambora, Indonesia	94	7
AD 1783	Laki, Iceland	134	6?
AD 969	Baekdu, China/North Korea	84?	7
AD 232 (1718 Cal. yr BP)	Taupo, New Zealand	270	7
AD 79 (1871 Cal. yr BP)	Vesuvius, Italy	95	5
3578 Cal. yr BP	Santorini, Greece	145	6
4260 Cal. yr BP	Hekla (H-4), Iceland	80	–
5150 Cal. yr BP	?Akutan, Alaska	175	–
5400 Cal. yr BP	?Towada, Japan	174	–

VEI: Volcanic Explosivity Index
Data from Zielinski *et al.* (1994), NOAA WDC-A, and other sources. Reproduced with permission of AAAS and Greg Zielinski.

Many secular climatic variations, in any case, do not appear to have been synchronous between different parts of the world (Wanner et al., 2008), and their recognition in palaeoenvironmental records is complicated by the fact that other natural agencies operated over similar timescales (see [Figure 1.1](#)). One such agency is wildfire, and in seasonally dry woodland ecosystems, the recovery time following fire disturbance and secondary succession is typically 50–200 years (Swain, 1973; Whitlock and Bartlein, 2004; Tinner et al., 2005). Fire affects the plant and animal community as a whole, but individual taxa react differently, some being harmed while others benefit. It creates landscape mosaics which often fail to achieve a full equilibrium with climate. Even so, combustible fuel load is one of the main factors to determine fire frequencies and intensities, and this in turn is strongly influenced by changes in climate. In regions that are relatively wet, such as western lower slopes of California's Sierra Nevada, fire scars in long-lived giant Sequoia trees have shown that wildfires increased during periods of drought (Swetnam, 1993). By contrast, in areas of soil-moisture deficit, such as the oak parkland zone of southwest Asia, combined analysis of lake charcoals and stable isotopes has shown that landscape burning was greatest during periods of wet climate, because here fires were fuel-limited (Turner et al., 2008). This means that while wildfire events often show a close relationship to climate change during the Holocene, there are significant differences between the fire history of regional biomes (Haberle and Ledru, 2001; Carcaillet et al., 2002; Power et al., 2008).

Another important agency in ecological change is disease pathogens, which, unlike fire, are host-specific. During the twentieth century, diseases have been responsible for the decline of several tree taxa, including the elm and the chestnut (*Castanea*). Chestnut blight was introduced into North America in 1904, and within 40 years, it had killed most of the mature trees in New England and the Appalachian mountains (Patterson and Backman, 1988). Dutch elm disease has been implicated in the mid-Holocene European elm decline (Girling and Greig, 1985; see Chapter 5), but it is difficult to prove a definitive link between this – or any other – former species decline and a disease outbreak. One of the most plausible examples is the reduction in hemlock (the conifer *Tsuga canadensis*, not the umbellifer herb that Socrates used to poison himself!), which took place throughout eastern North America at about 5.5 ka BP (Bhiry and Filion, 1996). Fine-interval pollen analysis of laminated sediments from Pout Pond in New Hampshire indicates that the decrease occurred within only seven or eight years ([Figure 6.3C](#)). This catastrophic decline appears to have been synchronous over the whole geographical and climatic range of eastern hemlock within the limits of ^{14}C dating. It was also followed by an increased concentration in the pollen of successional trees, such as birch (Allison et al., 1986). Hemlock subsequently remained rare for the following 1500 years, but eventually recovered, albeit much more gradually than its previous precipitate decline ([Figure 6.3A and B](#)).

Figure 6.3 A and B; Catastrophic mid-Holocene decline and gradual recovery of hemlock (*Tsuga*) trees in eastern North America, recorded in pollen records from two lakes (after Allison et al., 1986); C; Micro-stratigraphic pollen analysis from one of the lakes (Pout Pond) covering 350 years of the initial hemlock decline. Reproduced with permission of the Ecological Society of America.

A**B****C**

A problem already encountered with mammalian extinctions (see Chapter 3) and the elm decline (see Chapter 5) is that of ascertaining whether responsibility for specific environmental changes lay with human or natural factors. This problem becomes particularly acute during the mid- to -late Holocene. How, for

instance, should we interpret the changing Holocene concentration of greenhouse gases such as methane and carbon dioxide in the Earth's atmosphere prior to the Industrial Era (see [Technical Box VIII](#))?

The origin and development of blanket mires

A good regional-scale example of this problem of [equifinality](#) is the question of the origin of extensive areas of peat which today blanket the oceanic upland regions of the northwest Europe. On gently sloping plateaux peat is often raised above the general water table and receives mineral nutrients only from the atmosphere, mainly in rainfall. Most of these so-called [ombrotrophic](#) peat bogs were not present 10 000 or even 8000 years ago, but have developed during the mid- to -late Holocene as a result of some combination of climatic deterioration, soil retrogression and human impact (Charman, 2002; Simmons, 2003). Blanket peats have buried many upland soils and enabled new vegetation communities to become permanently established. For example, the blanket bog covering the high plateaux of Dartmoor, southwest England, will not revert to woodland even though relicts such as Wistman's Wood show that tree growth is still climatically possible here. The key factor in peat formation is waterlogging of the soil (see [Figure 6.5](#)). Waterlogging causes a reduction in cat ions and an increase in hydrogen ion availability (= fall in pH) which, with a suppression of microbial activity, fundamentally alters soil conditions. Only acid-tolerant plants such as ling heather (*Calluna vulgaris*) and cotton grass (*Eriophorum*) can survive in this environment, and the dominant component is usually different species of bog moss (*Sphagnum*). Like a sponge, *Sphagnum* is able to absorb water and hold it in the plant cells and so maintains its own water supply. As a result, a peat bog is always saturated with water.

The age and origin of blanket bog and other mires can be established by studies of peat stratigraphy, radiocarbon dating, pollen and plant macrofossil analyses. In some areas such as northwest Scotland, upland peat bogs can be regarded as climatically controlled ecosystems. Not only is there high effective rainfall here, but basal ^{14}C dates for peat initiation generally fall between 10 and 6 ka BP (Moore et al., 1984; Tipping, 2008), before the arrival of Neolithic agriculture. Elsewhere, on the other hand, many upland peats began to form at around the time of the mid-Holocene elm decline (Moore, 1975; Clymo, 1991), and they often contain a layer of charcoal at their base indicating that the previous woodland vegetation had been burnt. There is a strong suggestion that removal of tree cover triggered a sequence of environmental changes that led to woodland being permanently replaced by peatland (see [plate 6.4](#)). Trees, where they exist, fulfil a critical hydrological role by evapotranspiring at much higher rates than non-arboreal vegetation and by improving drainage via their rooting system. They also help maintain the circulation of nutrients. Trees can be killed not only by felling or burning but also by increased groundwater, and in ombrotrophic mires, groundwater levels are primarily related to the prevailing climatic conditions (Charman, 2002). Because the Atlantic/Sub-Boreal transition c.5.7 ka BP did not usually mark a shift to a wetter climate, a climatic trigger for blanket bog initiation at this time seems less likely than an anthropogenic one.

Once in existence, blanket mires evolved to a stable equilibrium state which nonetheless included periods of slower and more rapid peat growth (Blackford, 1993) (see [Figure 6.2](#)). In northern Scotland, tree stumps buried in blanket peat show that Scots pine was able to spread beyond its modern range for several centuries around 4500 years ago, probably because a period of warm climatic conditions dried out the peat surface (Gear and Huntley, 1991). Later, the Sub-Boreal/Sub-Atlantic transition around 2.6 ka BP saw a pronounced change in mire stratigraphy across northwest Europe indicating the opposite trend (Barber, 1993; Van Geel et al., 1996). Dark, oxidized peat typical of slow-growing mires and often including buried tree stumps was replaced by relatively undecomposed *Sphagnum* peat typical of wetter, fast-growing mires. This humification feature is often known as the [Grenzhorizont](#) (boundary horizon) after the terminology created by the German botanist Weber.

TECHNICAL BOX

Technical Box VIII: Ruddiman's early anthropogenic greenhouse hypothesis

Forest clearance and its conversion to farming and grazing land had local effects upon biodiversity, soil erosion, river flooding, and so forth. But did it – cumulatively – also have consequences for the global environment and climate? Bill Ruddiman (2003, 2005) has proposed that prehistoric and early historic land-use conversion led to an increase in the methane and carbon dioxide contents of the atmosphere thousands of years before the Industrial Revolution started. The ‘Ruddiman hypothesis’ began as a Sherlock Holmes-like detective enquiry, prompted by a piece of evidence that didn’t quite fit. Whereas atmospheric methane preserved in Antarctic ice cores peaked near the start of previous interglacials and then declined in concentration (see Chapter 3); during the Holocene it has steadily risen during the last 5000 years (Figure. 6.4). It is not hard to find natural explanations for the Holocene trend, such as boreal peatland emissions, but why would they have operated differently during the present interglacial compared to previous ones? The one thing that makes

the Holocene distinctive is human transformation of the natural world, in particular by farming and the associated change in land cover. Ruddiman proposed that the natural downward trajectories of key greenhouse gases has been reversed over the last five to eight millennia, primarily due to agricultural forest clearance for CO₂ and due to wet rice cultivation for CH₄ (Figure 6.4). These, he argues, may already have increased temperatures during the pre-industrial era by up to 1°C (and double that at high latitudes), and could even have prevented the onset of the next glaciation in northern Canada. During the late Holocene, the Antarctic ice core record includes a series of minor dips in CO₂ concentration, and Ruddiman argues that these relate to major pandemics, such as the sixth-century AD Justinian plague, which may have caused agricultural land abandonment and forest re-growth, with consequent carbon uptake from the atmosphere. The biggest of these carbon drawdown episodes probably occurred in the Americas after AD 1492, when European diseases killed >90% of the indigenous population and led to land abandonment and the growth of secondary forest (Nevle and Bird, 2008).

Is Ruddiman right? Debate has been lively, with one group of critics pointing out that new evidence from the long EPICA ice core (2004) indicates that MIS 11 – the closest interglacial analogue to the Holocene in terms of the Earth's orbital configuration – lasted up to 28 000 years, with CO₂ and CH₄ levels staying high, rather than declining progressively during the interglacial. Other critics include computer modellers who have calculated that the required land use-derived CO₂ emissions would have to be three to four times higher than originally estimated (Joos et al., 2004). Whatever the precise truth, the Ruddiman

hypothesis has forced scientists to reconsider the scale and antiquity of potential human impact upon the Earth system.

Figure 6.4 Expected and observed trajectories for atmospheric carbon dioxide and methane concentrations during the Holocene (modified from Ruddiman, 2005). Reproduced with permission of Princeton University Press.

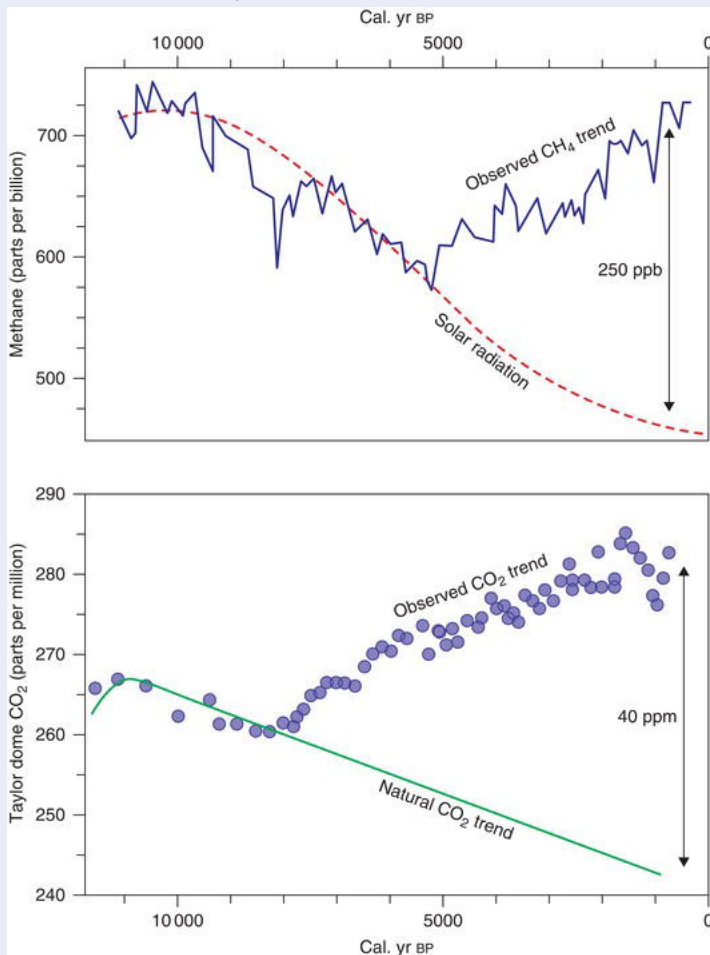
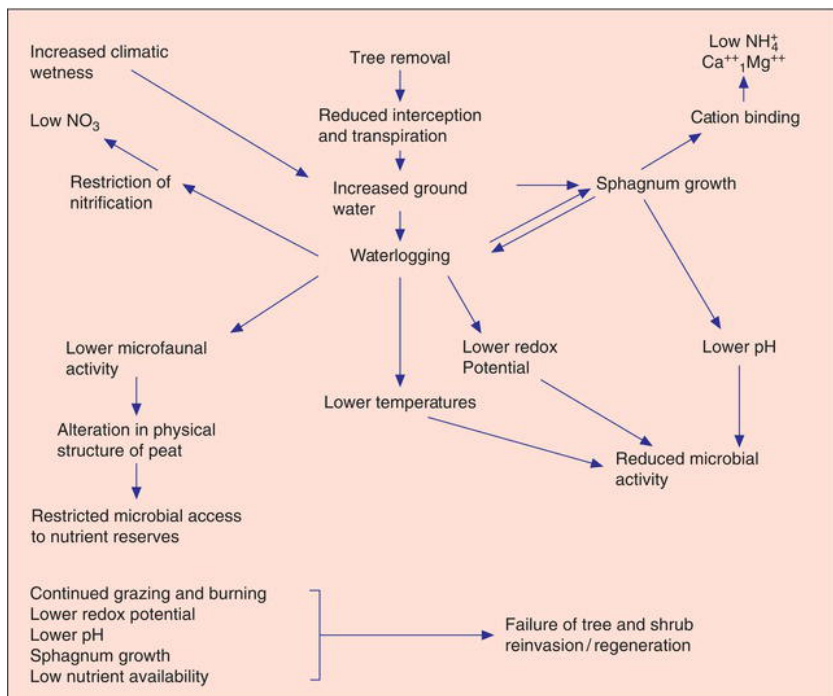


Figure 6.5 Model for upland peat initiation (Moore et al., 1984). Reproduced with permission of Elsevier.



Coasts and rivers

For most of the world's coastlines, the early Holocene was dominated by the continuing eustatic rise in sea levels as northern ice caps finally melted away (see Chapter 4). After around 7 ka BP, coastlines took on their familiar modern form, and with the exception of coasts experiencing glacio-isostatic rebound, the second half of the Holocene has been a period of relative sea-level stability (see [Figure 4.1B](#)). In detail, different parts of the world have each had their own local sea-level histories over the last 6000 years. In much of the Pacific, for example, sea levels lay several metres above their modern position during the mid-Holocene (Nunn, 1990; Mitrovica and Milne, 2002), whereas on the eastern seaboard of the United States, they have continued to rise gradually to their present elevations.

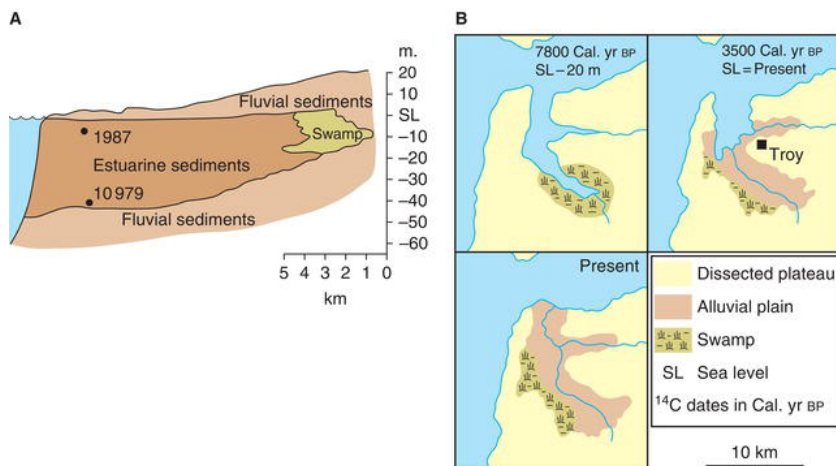
Plate 6.4 Tree stumps in peat cuttings testify to the existence of ancient forests at a time of dry climate on what is now a raised bog, Fallahogy, Ireland. Reproduced with permission of Jon Pilcher.



With sea-level stability came the formation of modern coastal landforms, such as Chesil beach in southern England and coral reef atolls in the Pacific. In estuaries and deltas, rivers started to reverse the long-lived marine incursion onto the land. As rivers deposited their sediment load, new land was built up and shorelines were pushed seawards again – a process which has continued up to the

present day (Strief, 2004). In the lower valley of the River Rhône, the sea reached its furthest landward position around 7000 years ago and has been in retreat ever since, as sedimentation has created new alluvial habitats including the Camargue (Pons et al., 1979). At the mouth of the Mississippi, no less than seven lobes-foot deltas have been built up and destroyed over the last 4600 years, while the Huang He and Yangtze rivers in China have pushed the coastline out by up to 250 km over the same period (Zong, 2004). A smaller scale but historically interesting sequence of early Holocene marine incursion and later Holocene progradation occurred in the Scamander river valley of northwest Turkey (Kraft et al., 1980). Boreholes show a wedge of sediments containing brackish-marine mollusc shells, ^{14}C dated to between c.11 and 2 ka BP, sandwiched between freshwater fluvial strata (see Figure 6.6). The Scamander valley was thus an embayment of the sea, and not an alluvial plain, at the time when the famous Bronze Age site of Troy (Hisarlik) was occupied. Homer's description of the landing of the Mycenaean Greek fleets during the Trojan War thus rests intact.

Figure 6.6 Geomorphic reconstructions in the vicinity of Troy during the Holocene (after Kraft et al., 1980). Reproduced with permission of the American Association for the Advancement of Science.



Inland from the coast, river systems were by now fully adjusted to regional climates and to the slopes, geology and natural vegetation cover in their catchments. Being dynamic systems, rivers periodically abandon their channels which has led to the creation of meander cut-offs where standing – rather than flowing – water conditions prevailed. These oxbow lakes underwent **hydroseral** succession, and many have become infilled with sediment – initially sands and gravels, later peats. The sediment archives of these old meander cut-offs are often rich in pollen and the remains of insects and micro-crustaceans (e.g. cladocera), which can provide information on changing floodplain ecology (Amoros and van Urk, 1989). From this, we know, for example, that most lowland English river floodplains were covered by dense alder woodland during the mid-Holocene (Brown, 1988). The dimensions of palaeo-channels and the size of the largest sediment particles at the base of their infill can be used to calculate their former

river discharge (Rotnicki, 1991; Bishop and Godley, 1994). Using maximum particle size data from the headwater streams of the Mississippi headwaters, Jim Knox (1993) has shown that relatively modest changes in the Holocene climate of the American Midwest led to large increases in flood magnitude. Because big floods only happen rarely, fluvial palaeohydrology provides one of the few sources of information on long-term flood frequencies and magnitudes (Kochel and Baker, 1982). Dating of Holocene river deposits suggests that climate shifts have acted as the main ‘pacemaker’ for the timing of major flood events (Macklin et al., 2005, 2010; Thorndycraft and Benito, 2006; Thorndycraft et al., 2012).

Cultural evolution

From a common starting point in peasant farming, there evolved during the second half of the Holocene a diverse range of cultures based on agriculture and/or pastoralism. The most important of these were socially stratified and involved Asiatic, Classical or feudal modes of production. All of these shared in common the production of an economic surplus by agricultural labour that was then taken and used by another social class. This new form of social relations also changed human–environment relations in several important ways. It meant, for example, that maximum economic output tended to replace the mix of social and economic goals that characterises communal peasant farming societies. Agricultural output was maximised by improving the productivity of existing farmland or by increasing the cultivated area. The latter led to clearance of woodland and draining of marshland, but also included cultivation of marginal land susceptible to soil erosion and other forms of damage. The advent of complex agricultural societies distanced and often weakened the link between people and nature. Nature became less the ‘habitat’ for the farmer than a set of economic resources to be managed and manipulated by the controlling group. This was particularly true of cultures where the dominant class was urban-based, as in Graeco-Roman antiquity. Although these complex societies shared certain key characteristics in common, such as urbanism and writing, they each had their own distinctive set of environmental relations. This is well illustrated by the Asiatic mode of production or, as Karl Wittfogel (1956, 1957) termed it, ‘hydraulic civilization’.

Hydraulic civilisation in Mesopotamia

The ancient civilisations of Egypt, Mesopotamia, South Asia and China all developed in a specific physical context – major alluvial river valleys under climates of low or unreliable rainfall. Wittfogel suggested that the agricultural potential of these fertile lands could only be realised by large-scale manipulation of their soil and water resources. This encouraged the development of hydraulic irrigation schemes, constructed under a centralised organisation. The agrarian economy in these great irrigated river valleys became the function of the state, which was also the sole owner of land. The timing and distribution of irrigation

water, the maintenance of canals and the collection and storage of surplus food were the responsibility of a professional bureaucratic ruling class which held power, typically despotic. The combination of organised human labour and a productive environment created a distinctive and potent mode of production which transformed nature so completely that almost nothing now remains of the original alluvial landscape in river valleys such as the Indus and Yangtze.

The environmental relations of hydraulic civilisation can be examined by analysing the transformation of one of the world's great river valleys, that of the Tigris–Euphrates, in what is today Iraq. The Tigris and the Euphrates rise to the north of lowland Mesopotamia in the mountains of Turkey and Iran. They receive winter rains which combine with spring snow melt to produce maximum river discharge in April and May. The winter crop is planted when the twin rivers are at their lowest levels, and the rivers then rise towards harvest time, 'threatening with inundation the crops they were with difficulty persuaded to germinate' (Ionnides, 1937, pp. 4–5). Irrigation agriculture was dependent on a 'good' flood: too high and settlements and grain stores would be ruined and too low and there would be a poor crop, food shortage and famine. In addition, there was the threat of river channels changing course. This occurred periodically as alluviation raised the height of distributaries, but could also receive human help when diversion canals were cut. Most of the water carried by the Tigris and Euphrates in fact never reaches the sea, but is lost as evapotranspiration from the flat alluvial marshlands. This brought another hazard to Mesopotamia, that of salinisation. As water is evapotranspired, river water solutes are left behind and eventually become concentrated to produce saline soils. These reduce crop yields and prevent cultivation altogether above salt levels of c.1%.

Simple floodwater farming involving use of residual soil moisture had been part of Neolithic agriculture in southwest Asia from the very beginning (see Chapter 5, and Sherratt, 1980). It is not surprising that the first archaeological evidence of irrigation canals appears as early as 8.2 ka BP, at Choga Mami on the slopes above Mesopotamian floodplain (Oates and Oates, 1976; Wilkinson, 2003). Occupation of the alluvial plain proper came in the protohistoric Ubaid and Uruk periods (7.3–5.7 ka BP) and lay along smaller distributaries of the two main rivers. These lay above the surrounding alluvial flood basins, and it was consequently possible to cut through the river levées to provide irrigation water for crops planted on the basin flanks. The natural channels were later straightened or replaced by completely artificial canals to build up a true irrigation network. Historically, Mesopotamian civilisation emerged into a dynastic phase around 5000 years ago (3000 BC) with monumental architecture, urban life and the world's earliest written records.

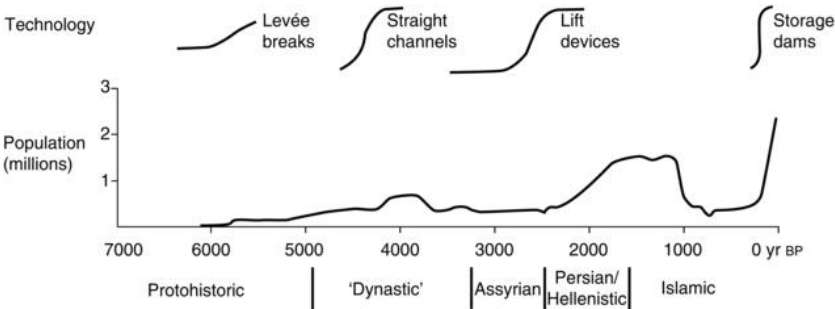
Despite its successes, irrigation agriculture in southern Mesopotamia encountered increasing ecological problems. Salinisation was kept at bay by alternating years of cultivation and weed fallow in which the leguminous perennials shok (*Prosopis farcta*) and agul (*Alhagi maurorum*) lowered the level of saline groundwater. This did not provide a permanent solution, and carbonised grains and textual sources show a switch between 4350 and 3650 Cal. yr BP (2400–1700 BC) from wheat to more salt-tolerant barley (Jacobsen, 1982). It was accompanied by a decline in barley yields and frequent mention in Sumerian texts of land abandonment due to salinisation (see [Table 6.2](#)). This agro-ecological

crisis was instrumental in bringing about a switch in the centre of power from southern to northern Mesopotamia, where saline soils were less prevalent. Irrigation in northern Mesopotamia needed to be larger and more complex, and the higher rate of siltation here meant that canals had to be regularly cleaned and maintained. On the Diyala floodplain north of Baghdad, over 10 m of alluvium has accumulated since irrigation began over 6000 years ago. Artificial regulation at nodal distribution points and control of the source of irrigation water led to a codified system of water regulation and centralised organisation, much as Wittfogel's model suggests. When this administration operated efficiently, for example, during the Sassanian and Abbasid periods (1750–1100 Cal. yr BP), agriculture flourished and population densities were high, but once central power was weakened, such as after the Mongol invasions (thirteenth century AD), there was land abandonment and disastrous population decline (see [Figure 6.7](#)). Archaeological surveys, notably in the Diyala region, clearly show this cyclical alternation between periods of stable irrigation and periods of disruption (Adams, 1965).

Table 6.2 Salinisation and changing cropping patterns in southern Mesopotamia
Source: Adapted from Jacobsen (1982)

	Cal. yr BP	Wheat (%)	Barley (%)	Ratio	Barley yield (l/ha)
Protohistoric	6000	46.5	53.5	1:1.5	–
Dynastic	4350	16.3	83.7	1:5	2537
	4250	3.0	97.0	1:32	–
	4050	1.9	98.1	1:53	1460
	3950	0.0	100.0	–	–
	3650	0.0	100.0	–	897
Early Islamic	1200	9.1	90.9	1:10	–

Figure 6.7 Historical changes in population and irrigation technology in lowland Mesopotamia (adapted from Bowden et al., 1981).



In Mesopotamia, there is a strong case for a direct causal link between hydraulic agriculture and the development of a centralised political state, but irrigation systems elsewhere in the world show this link much less clearly. In Egypt, for instance, nilotic agriculture was based on an essentially natural system

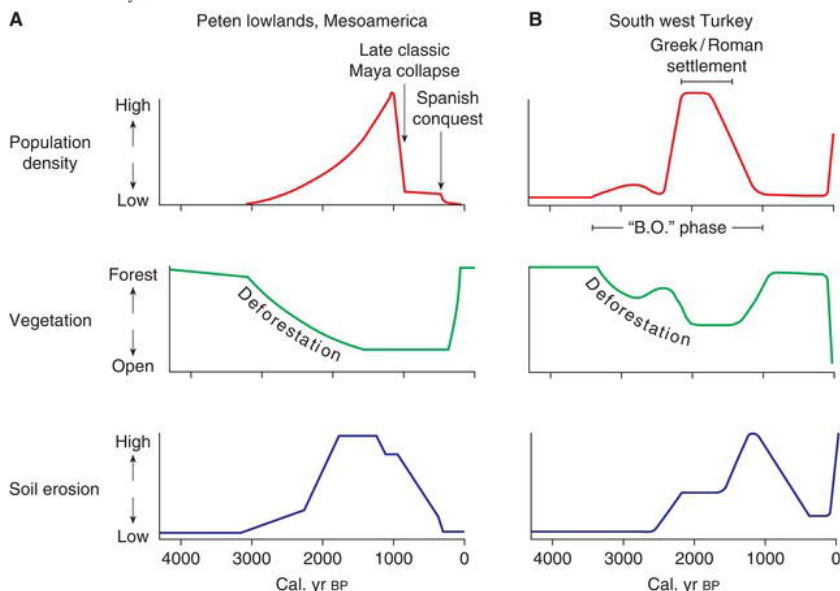
of flood-basin irrigation until only 150 years ago (Butzer, 1976). Unlike Mesopotamia, flood control and irrigation were locally administered, and there was a little need for a centralised bureaucratic apparatus. Competition for water between users was never more than a local issue in the Nile valley, and water rights did not have to be rigidly codified. In the Indus Valley, a major civilisation developed during the Bronze Age around urban centres such as Harappa and Mohenjo-Daro. Although clearly reliant on the waters of the river Indus, some of their most complex water management schemes developed as a response to a major drying of the climate c.4 ka BP (Staubwasser et al., 2003; MacDonald, 2011). This highlights the dynamic relationship between changing environmental conditions and complex agrarian and political systems. It is also important to recognise that while the natural environment may have stimulated economic and cultural developments, it did not pre-determine them. No great hydraulic civilisations developed on the floodplains of the Murray–Darling or Mississippi rivers. Equally, both natural forces, such as soil salinisation, and human ones, such as invasion from outside, could cause the complex ‘machine’ that was hydraulic irrigation to break down and collapse (Butzer, 2012).

Environmental impact in pre-Hispanic Mesoamerica

Complex, stratified societies took a little longer to develop in the New World than they did in the Old, but after 3000 Cal. yr BP (1000 BC), important indigenous civilisations did emerge in several centres within the Americas. Some of the most notable of these were in Mesoamerica, based on maize cultivation. One of the first great cultures to emerge with ceremonial centres and monumental architecture was that of the Maya, centred around the lowlands of the Yucatán Peninsula (Coe, 2002). The fact that ruined sites such as Tikal were cloaked in dense tropical forest when they were ‘re-discovered’ indicates the scale of Maya landscape clearance – which was more extensive than anything that has been achieved since. Dense rural populations were supported by a system of double cropping and in some areas used drained-field agriculture to exploit wetlands (Turner and Harrison, 1981; Darch, 1988; Furley et al., 1995). Population levels rose progressively from 1700 to 1100 Cal. yr BP, but this brought its toll on the predominantly limestone soils, which were subject to erosion and nutrient loss. Pollen records from lakes in the Peten region show widespread late Holocene deforestation (Deevey et al., 1979; Islebe et al., 1996; Leyden, 2002), and this was accompanied by a change in the composition of the lake sediments. Forest removal and urban construction were matched by increased accumulation rates of fine-grained mineral matter to create the ‘Maya clay’ found in many lakes (Binford et al., 1987; Brenner et al., 2002). Phosphorus was removed along with the topsoils, and phosphorus accumulation in the lakes closely tracks reconstructed population levels in the lake catchments (Figure 6.8A). The Maya’s increasingly precarious ecological support system appears to have been given a final push by a period of severe drought between 1200 and 1000 Cal. yr BP (750–950 AD) indicated by increased salinity levels in lakes (Hodell et al., 1995) and

sediment runoff events in Caribbean Sea varves (Haug et al., 2003). It has been argued that this environmental crisis, partly self-inflicted and associated with political instability and social unrest, led to the decline and, in some areas, the outright collapse of the Classical Maya civilisation (Gill, 2000). The last long count on a Maya jade ornament dates to AD 909; the peak of aridity in a nearby lake core lies at AD 922 (Hodell et al., 1995). In many areas, the ceremonial pyramids and other monuments were abandoned to the encroaching forest for a thousand years.

Figure 6.8 Late Holocene forest clearance and recovery linked to demographic change and soil erosion. A Maya lowlands (Binford et al., 1987, reproduced with permission of Elsevier); B southwest Turkey.



On the highlands of Mexico, drought was even more of a threat, and large-scale agriculture was often possible only by developing complex hydraulic irrigation schemes. The Classical period for Mesoamerican civilisation at sites like Teotihuacán was similar to that in the Mayan lowlands. It too went into decline after c.1100 Cal. yr BP, although the collapse was less sudden and less permanent. What is sure is that pre-Hispanic human impact was responsible for widespread landscape degradation. Sediment records from the basin of Mexico and in the volcanic highlands to the west record increased influx of eroded soil into lake basins, displaying regional differences in timing as different centres went through cycles of military success and defeat (Metcalf et al., 1989). In Lake Pátzcuaro, quantitative estimates indicate three periods of accelerated soil removal when erosion rates were at least as high as after the Spanish conquest, despite the absence of ploughs to turn the soil (O'Hara et al., 1993; Fisher, 2005). When Hernán Cortes crossed from the coast to enter the Aztec capital of Tenochtitlán in AD 1521, the Spaniard was encountering a far from pristine landscape.

Pastoral nomadism

Complex agricultural societies as existed in southwest Asia and Mesoamerica were not the only type of cultural development to emerge from communal peasant farming during the mid-Holocene. One of the most distinctive cultural developments was nomadic pastoralism, an extensive form of production adapted to arid or rugged terrain. Like hydraulic irrigation, pastoralism involves a distinctive set of ecological relationships between human groups and their environments, in this case mediated not through water use but through domestic animals. The domestication of animals such as cattle and sheep/goat had formed part of the early Holocene Neolithic revolution in southwest Asia, but a separate mode of production based on animal herding and involving a nomadic lifestyle only emerged during mid-Holocene times (Sherratt, 1981; Khazanov, 1984).

The domestication of the horse (*Equus caballus*) led to the emergence of pastoralist groups on the steppes of central Asia after 4000 years ago who were to be the forerunners of the later Huns, Turks and Mongols (Zvelebil, 1980; Anthony, 1991; Anthony and Brown, 2000; Outram et al., 2009). Nomadism developed around the same time in the newly desiccated deserts of Arabia and the Sahara where the camel (*Camelus*) was found to be better suited to the arid environment (Bulliet, 1975; Köhler-Rollefson, 1996; Wilkinson, 2003). In sub-Saharan Africa, pastoralists like the modern Masai were above all reliant on cattle, and forced southwards by increasing aridity, they spread progressively across that continent between 5 ka and 1 ka BP (Smith, 1980; Denbow and Wilmsen, 1986; Robertshaw, 1989; Marshall, 1990; Hanotte et al., 2002; Di Lernia, 2006). Like the plough, true pastoral nomadic societies such as these were restricted to the Old World prior to European expansion c.AD 1500, although a kind of agro-pastoralism based on llamas and alpacas did exist in the Andes during pre-Hispanic times (McGreevy, 1989). Pastoral nomads occupied land that was at best climatically marginal for settled farming and where the environment was inherently prone to great fluctuations in grazing, water and hence food supplies. Livestock herds not only provided foodstuffs such as milk and meat (Evershed et al., 2008) but also enabled pastoralists to move relatively swiftly to areas where grazing was seasonally available. Despite activities such as warfare designed to evade dependence on the environment, livestock numbers were traditionally held in check by the environmental constraint of animal carrying capacity.

Pastoral nomads have had a distinctive relationship with settled agricultural societies throughout the later Holocene. This has often been of a reciprocal kind involving trade and transport, for example, of salt in the Sahara. However, at other times, the relationship has been more hostile, with desert tribes attacking, overthrowing and replacing existing rulers of states and civilisations. In climatically marginal rain-fed farming areas of the Old World, the area of settled farming expanded and contracted as centralised political stability alternated with periods of weakness and the threat of nomadic incursions. In some cases, the stimulus to outmigration by pastoralists from their homelands was a prolonged run of dry years in which pasture for their herds shrank and pressure on resources increased. Ellsworth Huntington (1915) saw much of history as linked to the beating 'pulse of Asia', with climatic cycles pushing the Huns, Mongols and other

nomads outwards to the civilised rim of Eurasia. However, warfare is not the only possible response to animal carrying capacities being temporarily exceeded. At other times, pastoralists have simply sold off livestock to reduce the imbalance or settled down in adjacent land to become farmers.

Mediterranean ecosystems

Mediterranean ecosystems are among the world's most striking environments. Part of their fascination lies in aesthetic appeal, but it also has much to do with the complex and often long established pattern of human occupation and use of Mediterranean lands. Also, any environment which is capable of producing most of the world's great wines is surely worthy of attention! Climatically, it is possible to distinguish five main areas of the world with summer drought, winter rain of cyclonic origin and a mean annual temperature of $15 \pm 5^{\circ}\text{C}$. Central Chile, South Africa, South Australia, California and the Mediterranean basin itself all lie towards the western or southern margins of continental land masses at around 35° latitude. Their floras are physiognomically and ecologically similar, containing drought-tolerant forms such as the evergreen oaks of the Californian chaparral and the Mediterranean macchia. Mediterranean climates have produced broadly comparable plant formations, ranging from steppic grassland at one extreme to sub-humid forest at the other (see [Table 6.3](#)).

Of the major plant formations found in Mediterranean-type environments, only summer-dry evergreen forest, scrub and dry heath are distinctive to that ecotype. Sub-humid forest and steppic grassland exist elsewhere and cannot provide diagnostic evidence for the former existence of Mediterranean-type ecosystems. In the lands around the Mediterranean Sea, diagnostic taxa include evergreen oak, *Pistacia* and olive (*Olea europaea*), along with manna ash (*Fraxinus ornus*), hophornbeam (*Ostrya*), phillyrea (*Phillyrea latifolia*), box tree (*Buxus*), strawberry tree (*Arbutus unedo*) and some species of pine. Some of these plants have **sclerophyllous**, drought-resistant leaves which can be highly inflammable, so that Mediterranean forests are prone to fires, whether natural, accidental or intentional.

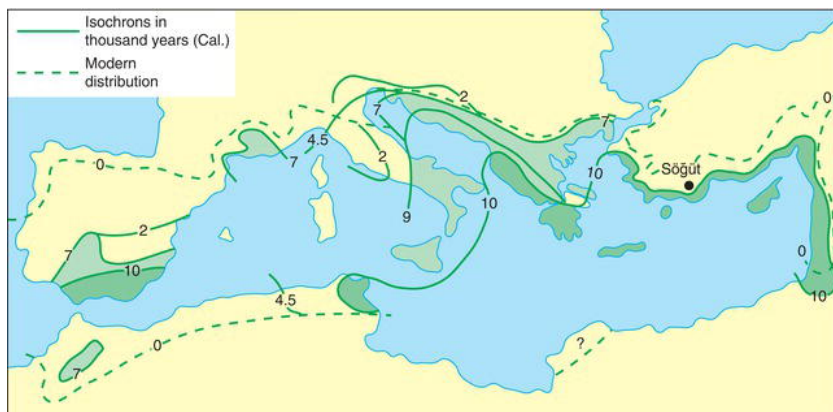
Table 6.3 Mediterranean-type plant formations

Source: Di Castri and Mooney (1973). Reproduced with permission of Springer Science + Business Media.

Formation	Life form	Height (m)	Understorey	Examples
Sub-humid forest	Broad-leaved deciduous or coniferous trees	5–> 30	Perennial grasses + herbs; sclerophyll shrubs on poorer soils	Montane forest (e.g. fir); deciduous oak chestnut
Evergreen forest	Broad-leaved evergreen trees (not all conifers)	5–30	Perennial grasses + herbs; sclerophyll shrubs on poorer soils	Forests of littoral pine, olive, carob, etc.
Open scrub	Stunted evergreen trees + sclerophyll shrubs	2–8	Perennial grasses, herbs + chenopods	Macchia, chaparral, matorral
Open scrub-heathland	Evergreen shrubs	0–2	–	Garriga, mallee heathland
Steppic grassland	Perennial grasses + herbs	0–1	–	Mediterranean steppe; tussock grassland

In addition to climate and vegetation, Mediterranean-type environments are also distinguished by their soils (e.g. red *Terra Rossa*) and by their stark, often memorable, landscapes. The image of bare white limestone and blue sea of an Aegean island comes easily to mind. However, landscapes are not only the product of geology and climate but also of human agencies, direct or otherwise. To what extent is that bare white limestone a product of anthropogenically induced soil erosion? What has been the role of fire in shaping Mediterranean plant communities? Indeed, how far are macchia and garriga natural vegetation formations at all, as they are associated so closely with human disturbance of the natural environment? According to Oliver Rackham's work in Greece, the characteristic Mediterranean scrub and heathlands are only maintained by cultural factors, such as grazing by sheep and goats (Grove and Rackham, 2001). If ecological succession were allowed to proceed rather than being retarded and held as a **plagioclimax**, macchia would evolve to become full woodland, while steppic grassland rather than heath would become the dominant non-tree vegetation. Pollen records from around the Mediterranean basin certainly show a marked increase in evergreen trees and shrubs during the later Holocene at the expense of sub-humid forest (see [Figure 6.9](#)), often linked to micro-charcoal evidence of more frequent wildfires (Vannière et al., 2011). How much the changed burning regime and vegetation cover were due to human impact and how much to climatic change is a subject much debated (Magri, 1995; Roberts et al., 2011).

Figure 6.9 Holocene spread of summer-dry woodland in the Mediterranean basin (data from Huntley and Birks, 1983, and other sources).

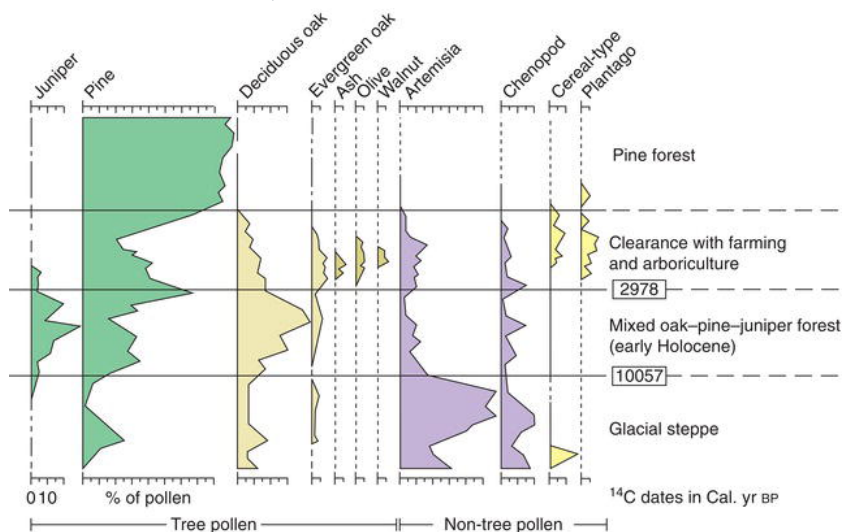


A comparison with the vegetation records in other parts of the world helps to clarify this issue (Roberts et al., 2001). Of the five Mediterranean-type regions, four have experienced major human modification only during the era of European colonisation (i.e. post AD 1500), but the fifth – the Mediterranean Sea basin itself – has a complex record of cultural–environmental relations stretching back to early Holocene times. This provides an opportunity to compare across time and space those Mediterranean ecosystems with short (<500 years) and long (>5000 years) records of agricultural impact. In fact, both California and southern Australia, although relatively free of human disturbance for the first 95% of the Holocene, nonetheless supported sclerophyll evergreen communities such as chaparral scrub and mallee heath over thousands of years before European contact, proving that these vegetation communities are not only the result of human disturbance linked to farming and grazing.

In the Mediterranean basin, evidence of early agricultural impact is hard to distinguish because Neolithic farming sometimes developed before ‘natural’ vegetation had stabilised after the last glaciation (Roberts, 2002). Once complex societies start to emerge around 5 ka BP, vegetation disturbance becomes clearly visible in pollen diagrams (Bottema and Woldring, 1990; Schwab et al., 2004). Societies such as Bronze Age Greece were associated with the development of the Mediterranean ‘triad’ of wheat, olive and vine, along with the cultivation of other trees such as walnut (*Juglans*) (Zohary and Spiegel-Roy, 1975; Bottema, 1980). Tree crops came to be domesticated not by seed selection but by other techniques of propagation, such as grafting. Unlike annual crops, cultivated trees have not extended much beyond their original climatic ranges, so that, for instance, olive groves remain a hallmark of Mediterranean landscapes. Arboriculture is a prominent feature of the so-called BO (Beyşehir Occupation) clearance phase which is present in pollen diagrams from Turkey (Vermoere et al., 2000; England et al., 2008; see Figure 6.10). In several pollen diagrams, this clearance phase starts just above a volcanic ash layer from the Bronze Age eruption of Thera (Eastwood et al., 1998) and lasted from around 3000 to c.1400 Cal. yr BP (1000 BC–AD 600). In southwest Turkey, the forest subsequently regenerated, but its species composition was changed. Prior to clearance, the forest had been a mixture of pine, cedar, fir and deciduous oak, while afterwards it was dominated

by pine alone. This may have been because forest grazing by livestock prevented re-growth of several tree species. Alternatively, it may be related to soil erosion during the clearance episode, with pine being able to survive better than other trees on the eroded, nutrient-poor soils (see Figure 6.8B).

Figure 6.10 Pollen diagram from Söğüt, southwest Turkey, showing Late Holocene clearance phase (data from Van Zeist et al., 1975).



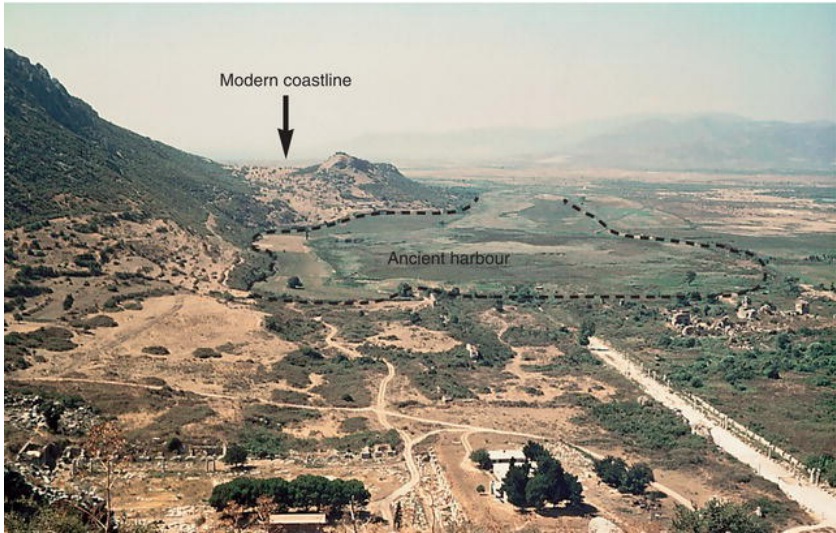
Palaeoecological records show that the intensity and timing of human impact on ‘natural’ forest vegetation varied from one part of the circum-Mediterranean region to another (Roberts et al., 2011; Sadori et al., 2011). For example, pollen analysis has shown that the magnificent cedar forests of the Middle Atlas mountains of Morocco continued unbroken right through the later Holocene (Lamb and van der Kaars, 1995). Human impact here came relatively late and until the nineteenth century was focused in Morocco’s lowland zone (Flower et al., 1989; Lamb et al., 1991; McNeill, 1992). In other areas, forests were modified and used as a managed resource for grazing, wood fuel and other products, rather than deforested; the *dehesa/montado* system for using Iberia’s oak woodlands, for example, has been shown by archaeology and pollen to go back to Bronze Age times (Stevenson and Harrison, 1992). Elsewhere in the Mediterranean, the late Holocene saw a major decline in fir forests, in some cases being replaced by beech and hornbeam (Reille et al., 1996; Watts et al., 1996; Magri, 1997). In the uplands of Corsica, beech forests were managed for pig pannage, but it is not always clear why human activities would have led to some of the changes that occurred in the species mix of Mediterranean forests, if indeed they did. It follows that deforestation was by no means the only way in which people transformed the native Mediterranean woodlands and that changes through time in the ratio of tree to non-tree pollen only tell us part of that story.

Mirroring the modifications made to natural vegetation, soil erosion has become widespread around the Mediterranean basin during recent millennia.

Evidence for this comes both from eroded hillsides and from valley alluviation. Many Mediterranean river valleys contain an alluvial deposit of historical age, which Claudio Vita-Finzi (1969) labelled the 'Younger Fill'. It contains pottery and other dateable remains which indicate that the main period of deposition was from late Roman to early medieval times (c.1700–500 Cal. yr BP). Mineral magnetic studies have also been able to trace its source to eroded catchment topsoils, rather than to stream incision of bedrock (James and Chester, 1995). But while the soil eroded from hillslopes helped to form new and fertile coastal plains, as at Troy (see [Figure 6.6](#)), it also choked the mouths of many rivers (Kraft et al., 1977; Bintliff, 2005). The classical harbour at Ephesus, for example, had to be abandoned because of siltation and now lies 4 km from the sea (see [Plate 6.5](#)) (Eisma, 1978).

Andrew Goudie's (2006) comment that 'in many cases of environmental change it is not possible to state without risk of contradiction that it is man rather than nature which is responsible' certainly applies to the origin of the Younger Fill. There has been particular debate about whether soil erosion and deposition were brought on by climatic change or by human land abuse (Wagstaff, 1981; Chester and James, 1991; Ballais, 1995; Provensal, 1995). The fact that alluviation began earlier in some areas than in others (Davidson, 1980; van Andel et al., 1990; Fuchs et al., 2004; Dusaar et al., 2011) suggests that climatic oscillations were unlikely to have been solely responsible. On the other hand, it would be wrong to assume that all eroded Mediterranean hillsides are anthropogenic landscapes. A salutary lesson in this comes from the badlands of southeast Spain. Here integrated geo-archaeological studies have found *in situ* Bronze Age sites lying un-eroded on vulnerable bare slopes (Wise et al., 1982). The badlands had already formed by c.4500 years ago, and later agricultural clearances did little to alter erosion rates. While giving the appearance of recent degradation, these and similar Mediterranean badlands, such as Cappadocia in Turkey (see [Plate 6.6](#)), are landscapes created as much by long-term geological instability as by the ravages of the goat and the plough.

Plate 6.5 Classical harbour of Ephesus, now 4 km from the sea due to historical river alluviation.



It seems that in most cases, historical soil erosion was a combined product of natural and cultural forces (Butzer, 2005; Woodward, 2009) so that attempts to explain them in terms of natural or human causes alone are probably misplaced. Mediterranean lands typically have thin or erodible soils, steep slopes, vegetation which is vulnerable to fire and rainfall that can be intense and erosive. These natural factors made Mediterranean ecosystems fragile under human impact. The replacement of forest by terraced vineyards and olive groves acted to conserve the soil, and so long as this agricultural system was maintained, land degradation was avoided. However, this was only a metastable equilibrium state. When the Roman Empire was invaded by barbarian tribes from the east, rural security declined and soil terraces fell into disrepair. The metastable equilibrium was upset, and in many cases, severe soil erosion followed, with valleys becoming choked by sediment. This helps to explain why the Younger Fill most often formed after and not during the Classical era of maximum land utilisation.

Plate 6.6 Ancient eroded badland, Cappadocia, Turkey.



Historical and palaeoecological research in the Mediterranean, as in many others areas, is increasingly based around integrated projects using a range of techniques – from pollen analysis to archaeological survey – to reconstruct regional landscape history (e.g. Jameson et al., 1994; Barker, 1995a,b; Zangger et al., 1997). These multi-disciplinary projects reveal that most Mediterranean landscapes are the result of human impact – direct and indirect – upon a fragile environment. The aromatic herbs, cork oak (*Quercus suber*) and bare white limestone that are considered to be the embodiment of the natural Mediterranean world have become dominant features only in late Holocene times. As the plagioclimax formations of macchia and garriga have increased, there has been a depletion of sub-humid forest, notably of deciduous oak, cedar and fir (Pons and Quezel, 1985). On eroded hillsides, it is doubtful whether these plant formations could ever return.

The making of the landscape: The British Isles

The landscapes of Britain and Ireland have long been a source of fascination, partly because of their diversity, but also because their development is so deeply rooted in history. These landscapes have been produced by the interaction of the peoples of the British Isles and their natural environments, now and in the past. But how is the history of that interaction to be reconstructed? One approach has involved using the landscape itself as a source of information (Aston, 1997; Muir, 2000). Celtic, Roman, Anglo-Saxon and Norman – these and other periods –

created their own cultural landscapes which were superimposed on pre-existing ones to create a palimpsest of elements of different age and origin (Hoskins, 1955). Distinctive features such as the feudal open field system of the English Midlands enable the landscape to be decoded and 'read like a book'. An account of the ecological history of the British Isles based upon elements that are still extant in the landscape might proceed as follows.

The natural, primaeval vegetation cover of lowland Britain was dense oak woodland except on sandy or calcareous soils where there was beechwood, grassy downland or lowland heath. By contrast, much of highland Britain and Ireland was covered by moorland or blanket bog. These highland wildscapes have undergone much less cultural modification than lowland landscapes, at least up until twentieth-century coniferization. The main clearance of native woodland occurred in the Roman or in the early Medieval periods. Prehistoric peoples, though leaving their imprint on the landscape in the form of stone circles and barrows, were few in number and their impact small in comparison with historical times.

In fact, most if not all of an account such as this would be incorrect. This is because landscape-based reconstructions are able to cope much better with recent than with distant time periods. Very little of those earlier prehistoric and primaeval landscapes escaped the shipwreck of time, and the few surviving relicts are not always representative of what was formerly widespread. Because information shrinks rapidly backwards through time, historical myopia is created – our vision is clear for things that are close in time but becomes blurred for those in the more distant past.

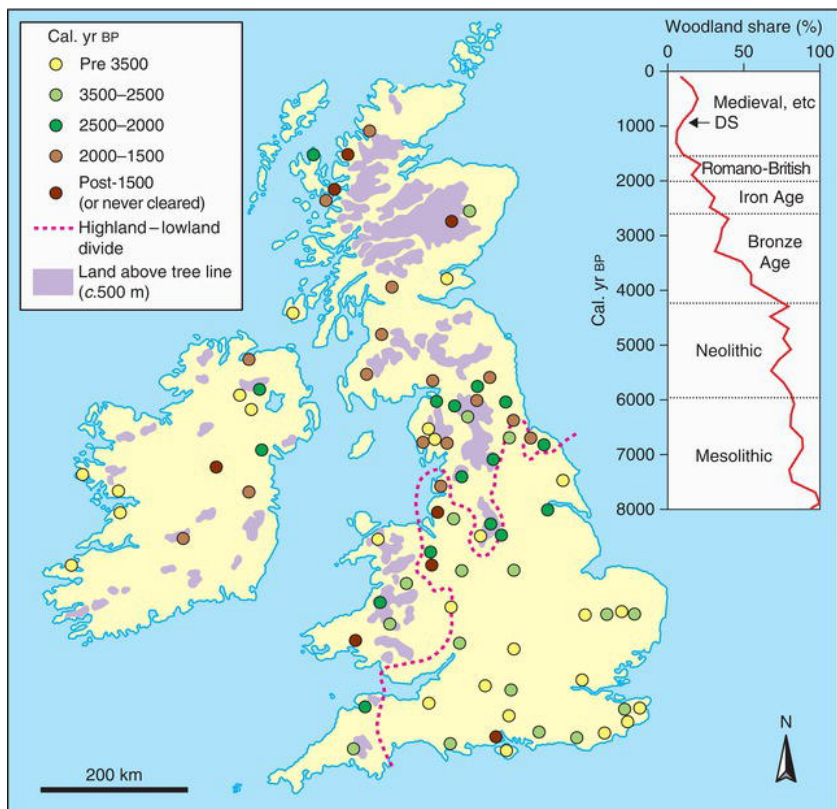
The same is also true of a second source of information – documentary evidence. How, for instance, can the medieval Norman and Iron Age Celtic contributions to landscape change be realistically compared, when one is well documented historically and still evident in the landscape while for the other there are no written records and whose physical evidence has largely been reduced to hilltop ditches and embankments? Fortunately, one documentary source does provide an extremely important baseline datum, for the lowland zone of England at least, from which landscape studies can work backwards. That source is the Domesday survey of AD 1086. King William's survey of his newly conquered kingdom tells us that a little over nine centuries ago lowland England was already densely peopled, with the Weald and the Forest of Dean as two of the few remaining areas of wildwood. Over most counties of eastern England and also in Devon and Cornwall, woodland accounted for less than 5% of the total land area. Oliver Rackham (2003) has estimated that only 15% of Domesday England as a whole was wooded, less than half the area then under arable land. Although the Normans were responsible for reducing the woodland cover still further, from 15 to 10% by AD 1350, it is clear that the vast majority of primary forest clearance had taken place before the Norman conquest of England in the eleventh century AD. If we are to establish when and how the 'natural' landscape of around 6000 years ago came to be the largely cultural one recorded in the *Domesday Book*, then non-documentary evidence from archaeology and palaeoecology becomes critical (Jones, 1986; Simmons, 2001).

The primaeval forest

The notion of an original pristine British landscape or *urlandschaft* is something of a myth, because even in the early Holocene, Mesolithic hunter-fisher-foragers were capable of significant ecological impacts (see Chapter 4). The nearest equivalent to a primaeval, naturally vegetated landscape would be the forests of the Atlantic period, between about 8000 and 6000 years ago. Pollen diagrams show that at this time the only substantial areas without forest were the 5% or so of the British Isles above the tree line which lay at around 500 m (see [Figure 6.11](#)). At that time, there were trees on what today are bare and windswept landscapes like the remote Shetland Islands to the north of Scotland (Bennett et al., 1992). Even many coastal wetlands supported trees, as is shown by submerged beds of pine, oak and yew (*Taxus*) uncovered from Fenland peat (Godwin, 1978).

The composition of the forest varied significantly across the country, forming a polyclimax vegetation (Bennett, 1989). Over most of the highland zone, woodland was dominated by the sessile oak (*Quercus petraea*). In Scotland around the Great Glen lay the Caledonian pine forest, a relict of the pinewoods which had briefly covered most of the British Isles early in the Holocene. Elsewhere in the highland zone, local variants of the flora included birchwoods in the far north of Scotland and elm woods in the middle third of Ireland. Blanket bog had started to develop in some parts of Ireland and Wales but still only formed a minor component of the landscape. The character of lowland zone woodland was equally varied, although those variations were generally more subtle in the way they merged with each other. Despite the oft-quoted ‘mixed oak forest’ label, oak was by no means the dominant tree species. In many parts of lowland England, such as East Anglia, limewoods (*Tilia*) were extensive (Greig, 1982), while alder (*Alnus*) occupied the damp ground in river floodplains and around lakes (Brown, 1988). Hazel (*Corylus*) was a widespread understorey shrub throughout the British Isles, and locally it formed substantial trees. Beech, although present, did not become an important component of woodland until the Bronze Age.

Figure 6.11 Map showing dates of first major forest clearance phase in Britain and Ireland from pollen and land snail evidence (data from Turner, 1981, and other sources). Note the generally early date of initial clearance across lowland England and the more varied ages in the highland zone. Inset shows changes in woodland cover for Britain inferred from pollen data (data from Woodbridge et al., 2013). DS marks the time of the Domesday survey.

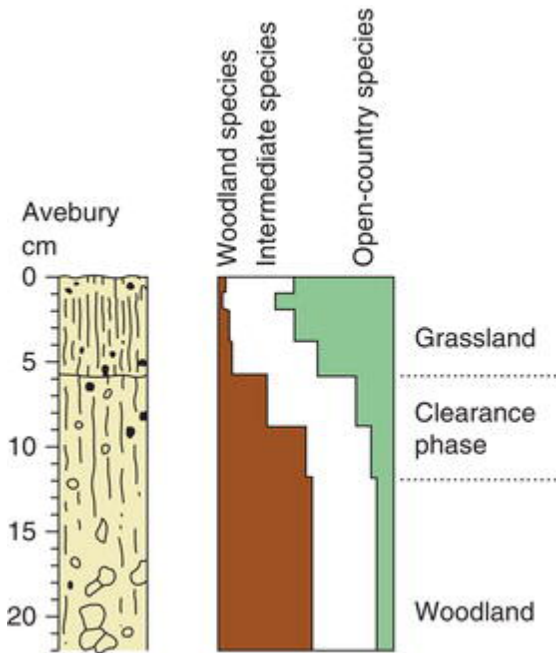


By the time Neolithic agriculture reached the British Isles around 6 ka BP (Brown, 2007), eustatic sea levels were within 2 m of the present day and the coastline had its familiar modern configuration. Consequently, the first farmers had also to be familiar with sailing in order to cross the English Channel and the Irish Sea with their new domestic crops and livestock (Rowley-Conwy, 2004, 2011; Collard et al., 2010). As elsewhere in northwest Europe (see Chapter 5), the initial impact of agriculture was relatively slight, clearances of the forest being localised and temporary. Forest fallow agriculture, where it operated, created a mosaic of secondary woodland in various stages of regeneration. The forest cover was therefore largely maintained, but its species composition altered; elm, for instance, declining in importance after around 5.7 ka BP (Parker et al., 2002). At a landscape scale, therefore, the early Neolithic farmers modified the pre-existing woodland ecosystems rather than replacing them completely by agro-ecosystems. Rather than continuing to encroach progressively into forests, the impact of agriculture decreased in the middle Neolithic (5.3–4.5 ka BP) when there was a widespread and extended phase of woodland regeneration (Whittle, 1996; Stevens and Fuller, 2012; Woodbridge et al., 2013).

Neolithic and early Bronze Age society reached its apogee between 4.5 and 3.8 ka BP with the completion of great monuments at Silbury Hill and Stonehenge. These are not only impressive feats of engineering but also testimony to the

existence of periods of labour surplus within the prehistoric economy. The main legacy in the landscape of Britain's first farmers is consequently not their farms or villages (a part of the economic process) but their burial and ceremonial places (a part of the social process). The obvious importance of social activities should make us cautious in explaining Neolithic or early Bronze Age impact on flora or fauna solely in terms of agricultural economics. For example, land snails and pollen from buried soils at Avebury, Silbury Hill and Stonehenge show that the chalk landscape had by then already been changed from woodland to open pasture or scrub (see [Figure 6.12](#)). The area around these sites may have been kept clear by grazing animals so that the monuments remained visible for astronomical observations or other non-economic reasons. Evidence of permanent pasture around ceremonial and burial sites need not mean that, for example, agricultural land was not being periodically abandoned and reoccupied elsewhere in the landscape.

Figure 6.12 Land snail record of woodland clearance, Avebury, southern England (Evans, et al. 1972).



The period 3.8–2.6 ka BP saw the end of a monument-building society with its roots in Neolithic times and the beginning of what would become Celtic Britain (Hunter and Ralston, 1999; Pryor, 2003; Stevens and Fuller, 2012). From both social and environmental points of view, this shift was fundamental. Starting in the Middle Bronze Age, society developed by the Iron Age to become a hierarchical, tribal one, and its mode of production became progressively less dependent on domestic subsistence agriculture. The associated demographic changes were equally profound. Britain's population is estimated to have risen

five- or even 10-fold between 4000 and 2000 years ago from perhaps 200 000 in the early Bronze Age to as much as two million by the late Iron Age (Pryor, 2003), the latter figure being higher than that recorded in the Domesday survey. These changes were manifest in the landscape in the creation of organised arable field systems and other forms of land allotment and defensive settlements such as 'hill forts' (Bowen and Fowler, 1978). Although ard marks are known from Neolithic contexts, it was after 3.6 ka BP that the use of the ard (or scratch plough) appears to have become widespread. Animals were no longer raised solely for meat, but also came to provide traction for ploughing, manure to maintain arable soil fertility, transport, wool and milk (Sherratt, 1981). New crop strains were also adopted, notably spelt wheat (*Triticum spelta*) and the Celtic bean (*Vicia faba*). In other words, after an era of social investment in ceremonial monuments, spare productive capacity was switched to the more pragmatic task of creating Britain's first large-scale agricultural landscapes (Fowler, 1983, p. 36). The period from 3.8 to 2.6 ka BP also saw an important extension of settlement in many parts of highland Britain. A good example is provided by Dartmoor where large parts of the Bronze Age landscape have been preserved (Price, 1993).

Shaugh Moor: A Bronze Age landscape

Southwest Dartmoor is today an area of china clay (kaolin) workings, and during the 1970s and 1980s, a rescue project took place to investigate a 5 km² block of landscape before it was destroyed by the bulldozer. The Shaugh Moor project from the outset involved not only archaeological excavation and survey but also studies of soil and vegetation histories, in order to establish the natural environment of the Bronze Age. The integrated approach employed at Shaugh Moor allowed local spatial variations in environmental conditions to be tied directly to patterns of past human land use.

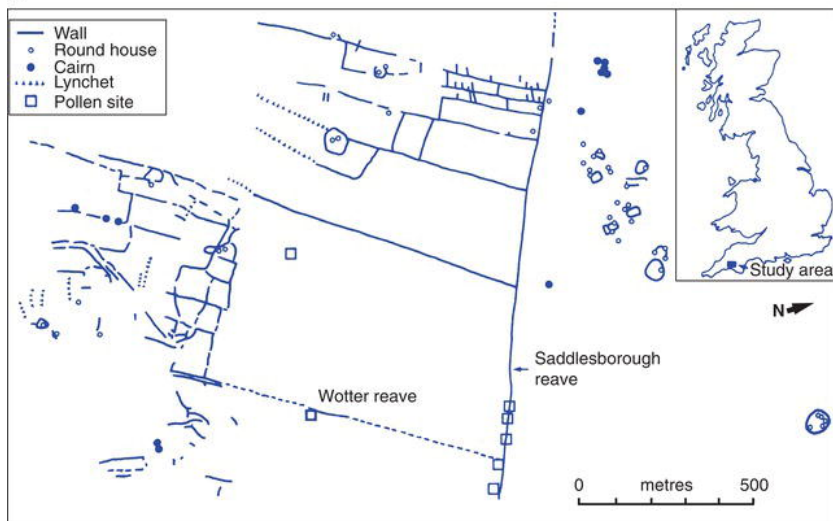
Shaugh Moor incorporates a series of low stone walls – or reaves – some of which are linked to the wider system of prehistoric land boundaries that covers the whole of the Dartmoor uplands (Fleming, 2008), while others form smaller areas of enclosed pasture (see [Figure 6.13A](#)). Although animal bones were not preserved under the acid soil conditions, hoof prints of sheep and cattle were found in a Bronze Age reave ditch, and pollen of pastoral weeds was much more abundant than that of arable species. Carbonised cereal grains were virtually absent, and phosphate levels do not indicate manuring within the enclosed fields. The economy was therefore evidently based on pastoralism rather than arable farming. Settlement evidence included round, stone-built farm dwellings which may have only been occupied seasonally, perhaps as part of a transhumant pattern of land occupation. Both dwellings and reaves involved extensive use of wooden posts, indicating the availability of timber and its exploitation. ¹⁴C dates on wood charcoal and peat show that the reaves were built around 3.8 ka BP and that they and the settlements were used for about 1200 years.

Pollen samples were studied from peat deposits and soils below and above Bronze Age reaves. These and other Dartmoor pollen diagrams (e.g. Fyfe et al., 2008) show evidence of Neolithic clearance and subsequent woodland

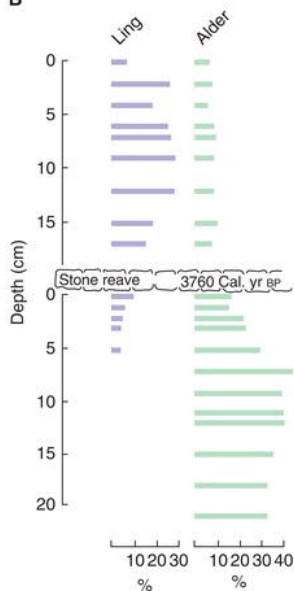
regeneration prior to reave construction. As archaeological remains of Neolithic Age are hardly recorded at Shaugh Moor – probably because of their subsequent destruction in the Bronze Age – this pollen evidence serves as a reminder of how selective cultural landscape preservation can be. By 3.8 ka BP, the vegetation was a mosaic of alder woods, hazel scrub on the better drained slopes and acid grassland on the plateau top above about 300 m. After a millennium of Bronze Age pastoral occupation, most of this had been transformed into moorland, with ling (*Calluna*), other heathers and acid grasses growing on peaty stagnopodzols (see [Figure 6.13B](#)). Soil changes from acid brown earths to podzols and gley soils had already commenced when the Bronze Age occupation began, but under it, degradation continued so that by 2.6 ka BP (650 BC) both soils and vegetation were indistinguishable from those of today. Shaugh Moor seems to have been virtually abandoned at this time, leaving behind a ‘fossilized’ Bronze Age landscape disturbed only by subsequent mining for tin and china clay.

Figure 6.13 Shaugh Moor, Devon, England; A relict Bronze Age cultural landscape (based on Smith et al., 1981); B pollen diagram from above and below Wotter reave (data from Balaam et al., 1982). Reproduced with permission from the Prehistoric Society.

A



B



The environmental impact of permanent agricultural clearance

It has been argued (Burgess, 1985) that climatic deterioration, notably at the Sub-

Boreal/Sub-Atlantic transition around 2.6 ka BP, was critical to the human ecology of Britain's highland zone. The Bronze Age had marked the 'high tide' of upland farming in many parts of the British Isles, and at least in some areas, the population subsequently retreated to lower ground in the face of an increasingly hostile climate, deteriorating soil and vegetation resources and a decline in demand for mineral resources such as tin. On the other hand, pollen evidence suggests that this process varied in its timing and severity between different upland regions (Tipping, 2002; Dark, 2006; Amesbury et al., 2008). Nonetheless, it seems that environmental changes which prompted land abandonment in Britain's uplands could be as permanent and damaging to human welfare as the desertification that has affected parts of Africa in recent decades. And like African desertification, these changes were due at least partly to human misuse of a fragile environment. Dartmoor's prehistoric pastoralists had almost literally been treading on dangerous ground.

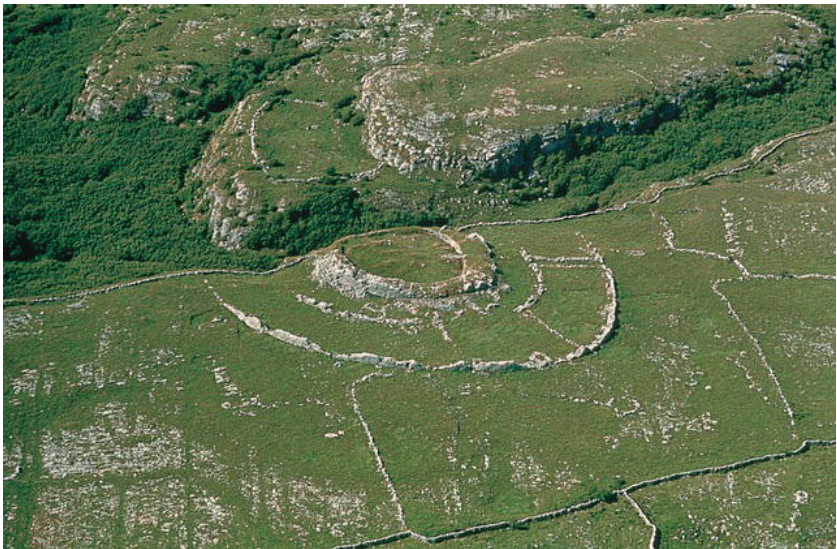
If the Bronze Age 'enclosure movement' created a landscape of environmental dereliction in parts of northern and western Britain, it had precisely the opposite effect over much of lowland England. Here, landscape reorganisation was associated with an increase in agricultural productivity and carrying capacities. The small scale of Neolithic cultivation allowed most of the woodland cover to be maintained, but with 'enclosure' came permanent clearance of the forests of the lowland zone. Quantitative pollen-based estimates of the regional extent of forest clearance (Woodbridge et al., 2013) show that the Bronze and pre-Roman Iron Ages (i.e. 4–2 ka BP) were the most active period of wildwood destruction in the history of Britain (Figure 6.11).

From Julius Caesar's description that 'the population [of southern Britain] is exceedingly large, the ground thickly studded with homesteads' (*De Bello Gallico*, Volume 2), it certainly appears that the Romans took over a landscape in lowland Britain that was already largely agricultural. This is confirmed from aerial photographs which show dense Iron Age settlement, for instance, on the gravel terraces and floodplains of major rivers such as the Thames (Jones, 1986). At the northern margin of their empire, however, the Romans may have contributed to woodland clearance. The construction of Hadrian's Wall – to keep out the Caledonian Picts and Scots – used large numbers of mature oak trees (McCarthy, 1995), and pollen diagrams from peat bogs along the wall show a dramatic change from a mainly wooded to an open landscape at about this time (Dumayne and Barber, 1994; Manning et al., 1997). Despite their different social bases, Celtic, Classical and early feudal modes of production in the British Isles were all overwhelmingly agrarian in character (Pryor, 2003). The post-Roman period witnessed a decline in population and woodland regeneration in some areas of England (Oldfield et al., 2003), but the basic pattern of land occupation established in the pre-Roman Iron Age was not greatly altered. It seems that the distribution of wooded and non-wooded land and the territorial divisions of most estates and parishes recorded in Domesday England had already then been in existence for at least a thousand years, and possibly twice that long (Pryor, 2006). In Ireland, forest clearance followed a similar pattern to that in Britain during later prehistoric times, but with a significant increase in archaeological settlement density after ~1600 Cal. yr BP (~400 AD) with the construction of cashels and other circular enclosures (see Plate 6.7). This is the time of the oldest Irish annals

and other documentary records, many of them linked to the arrival of Christianity (Mitchell and Ryan, 2001).

In medieval times, the survival or destruction of English woodland lay almost entirely in the hands of the dominant social classes – the Church, monarchy and landed aristocracy – notably as royal forests and deer parks. In contrast to prehistoric and early historic times, there was rarely common access to woodland resources. Woodland had formerly been less valuable than other forms of land use, but after AD 1250, the value of trees and wood fuel rose steadily, providing landowners with a good income as well as the pleasures of the hunt (Rackham, 2003). Woodland conservation and management made economic sense, and forests were ruthlessly protected against ‘wood stealers’. Damage to hollies or thorns – which offered safety to young oak trees – could lead to three months of hard labour with a whipping in each month, for example. On the other hand, the conservation ethic did not apply equally to rich and poor. In 1251, King Henry III’s Christmas dinner included 730 deer, 1300 hares, 395 swans, 115 cranes and 200 wild boar – the boar coming from the last remaining population in England which became extinct soon afterwards (Rackham, 2003).

Plate 6.7 Former field systems on the Burren, western Ireland, now containing limestone pavement and largely devoid of soil. This cashel (defensive round enclosure) at Cahercommaun was occupied from the fifth to the tenth centuries AD; behind are the remnants of woodland which was formerly extensive in the area. Photo credit: University of Cambridge, Committee for Aerial Photography.



Pollen records tell an essentially similar history of land use to those from archaeology and documented history, with woodland cover across Britain as a whole declining progressively between 4000 and 1000 years ago (Figure 6.11). This landscape transformation can be illustrated by a pollen diagram from Rimsmoor in Dorset which, in many ways, serves to summarise the whole history of human impact on Britain’s lowland zone vegetation during the Holocene.

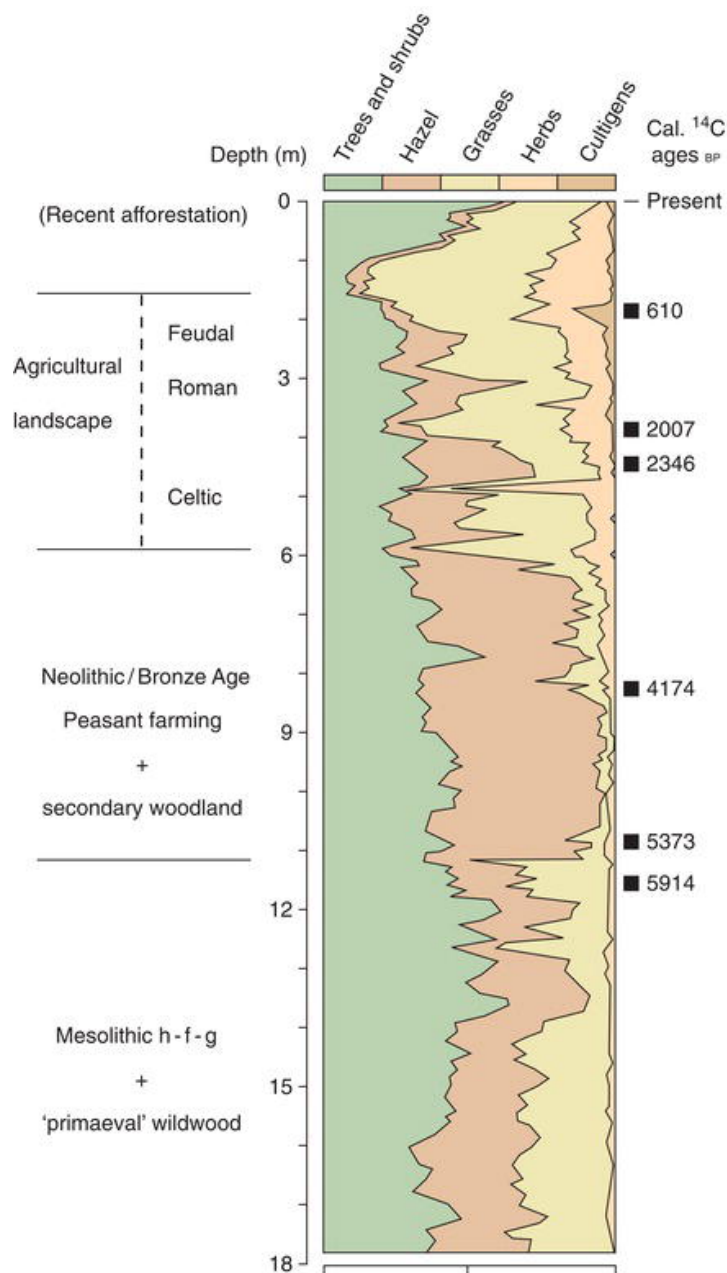
Rimsmoor is a small bog lying near to the edge of the chalk and contains no less than 18 m of pollen-bearing sediment. As a result, it contains a detailed record of mid- to -late Holocene ecological change, with a pollen count every three to five years for key sections of the diagram (Watson, 1982). Prior to 5.7 ka BP, the vegetation comprised a stable mixed deciduous woodland community, with oak, elm, ash, lime and alder as the main tree taxa (see [Figure 6.14](#)). Abundant charcoal indicates periodic Mesolithic burning of the landscape. Between 5.7 and 3.2 ka BP, hazel was abundant, oak fluctuating, herbs and grasses gradually increasing, with cereal pollen occasionally present. This is what might be expected under a system of extensive forest fallow agriculture, with much of the land under regenerating secondary woodland or scrub. Modification of woodland species composition is indicated by a decline in elm (c.5.7 ka BP) and lime (c.4.2 ka BP).

Clearance associated with early 'Celtic' landscape organisation around 3000 years ago is very clearly marked in the Rimsmoor pollen diagram. There is a sharp increase in herbs and grasses including those, such as *Rumex* and ribwort plantain (*Plantago lanceolata*), typical of agricultural activity. Over the next two millennia, the intensity of land use appears to have been broadly constant, despite periodic changes in the balance between pastoral and arable components of the landscape and the introduction of some new crops, such as rye (*Secale*) and hemp/hop (*Cannabis*) (see also Edwards and Whittington, 1992). Cultivation reached a peak of intensity in medieval times around the fourteenth century AD and subsequently declined, with re-forestation by pine in the eighteenth and nineteenth centuries. In short, this pollen diagram records three phases of human land-use activity prior to the advent of modern times, each one lasting about 2500 years. The first was hunting and gathering under wildwood; the second, small-scale peasant farming within secondary woodland; and the third phase dominated by an agricultural landscape of fields and farms created under complex, stratified societies.

The later Holocene woodland clearance recorded at Rimsmoor and many other pollen sites had permanent consequences for catchment soils. The original brown earths have in some cases been preserved beneath Neolithic and Bronze Age barrows and earthworks, and they are quite different from the soils which now surround them on the chalklands. In particular, the fertile but superficial cover of loess has been almost completely eroded to leave the present thin calcareous soils (Catt, 1978). The eroded soil has 'sludged' downhill to form extensive colluvial deposits at the base of slopes (Bell, 1983). Some of this was moved into river systems, whose sediment load was increased further by clearance of alder woodland on lowland floodplains (Brown, 1988). These influxes of sediment led to the widespread accretion of minerogenic lake sediment and fine-grained floodplain alluvium after c.3000 Cal. yr BP (Brown, 1997; Edwards and Whittington, 2001), while upland rivers underwent more complex cycles of river incision and aggradation (Macklin et al., 1992). While the timing of major river flooding events appears to correlate best with phases of climatic deterioration (Johnston et al., 2006), their overall impact was significantly increased by the long-term loss of forest cover. On the chalkland, erosion of loess was not normally sufficient to prevent tree regeneration. But in other environments, pedogenic changes had far-reaching consequences for the total

ecosystem which prevented it from reverting to its original state. One such extreme example of irreversible ecological change is provided by the limestone plateau of the Burren in western Ireland. The Burren's thin soil cover, which was nonetheless able to support pine, yew and birchwoods during the Holocene, has been almost totally eroded down karstic fissures during the recent millennia (Drew, 1982; Watts, 1984; Feaser and O'Connell, 2009). All that is left is bare limestone pavement incongruously criss-crossed by Celtic fields with no soil inside them (see [Plate 6.7](#))!

Figure 6.14 Pollen diagram from Rismoor, southern England (Watson, 1982; reproduced with permission of Paul Watson).



Conclusion

It is easy to fall into the trap of describing human impact on the natural world solely in terms of ‘degradation’ and ‘impoverishment’, especially when considering issues such as soil erosion or deforestation. In fact, agriculture – at least in its pre-modern form – has generally been an agent of ecological diversification. It caused the relative homogeneity of primeval forest ecosystems to be replaced by a mosaic of habitats, from downland to heath, macchia to garriga. These are all semi-natural **plagioclimax** ecosystems, created and maintained by human action, and their fates came to be intimately associated with particular modes of agricultural production. The English chalk downlands, for example, were maintained as a grassland ecosystem only because of a long-standing system of sheep grazing. Once this traditional form of land use no longer operated, the natural processes of plant succession began to allow annual grasses to be replaced by scrub and eventually to revert to secondary woodland. The ‘ecological clock’ was in this case held up by another domestic herbivore – the rabbit – but this proved a temporary check when myxomatosis decimated the rabbit population in the 1950s. Consequently, conservation of the springy turf and panoramic views of England’s downs will not be achieved by simply protecting that environment from human disturbance and leaving nature alone.

The long interplay between human production and different natural environments often makes it effectively impossible to separate natural from cultural influences (Head, 2000; Oldfield, 2008). All landscapes created by agricultural societies are intrinsically both natural and cultural. As a consequence, it can be difficult to pinpoint simple cause-and-effect mechanisms for environmental changes such as soil erosion or blanket bog formation (Bell and Walker, 2005). In practice, it would be artificial and illusory to separate out what was the result of *interaction* between people and nature (Head, 2008). The environmental changes that occurred during the period from the mid-Holocene until the last millennium cannot be neatly and succinctly summarised; they were too diverse through time and across space. However, diversity was to give way to uniformity in more recent times, not due to natural mechanisms but to human ones. The creation of the European world system and industrial capitalism and their impact on the natural world form the focus of the next chapter in this story.

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The impact of modern times (1000–0 Cal. yr bp)

Introduction

The last millennium of the Earth's natural history has been a time of dramatic and accelerating change. One has to look to the beginning of the Holocene, with the climatic amelioration after the Last Ice Age and the Neolithic agricultural revolution, to find a period which produced changes of comparable significance for human–environment relations across the planet. What makes the last few centuries so distinctive is the quantum leap in human impact on nature that has occurred during this time. Climate and other natural agencies have been far from static over the last thousand years, but there are no grounds for believing that they operated in any way that was fundamentally different from earlier in the Holocene except where prompted by human agency. The justification for treating the period since AD 1000 as a separate timeblock within the Holocene is therefore based on culturally induced changes in the natural environment. The exponential increase in human impact on the global environment even led Paul Crutzen and Gene Stoermer to propose that the last few decades or centuries should be designated as a new geological epoch, which they labelled the **Anthropocene** (see [Technical Box IX](#)).

TECHNICAL BOX

Technical Box IX: The Anthropocene

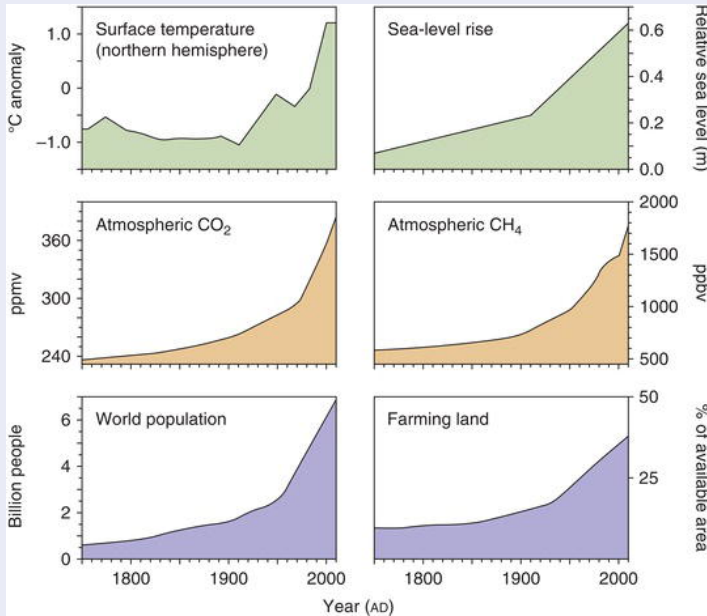
Have we entered a new geological epoch? Paul Crutzen and Gene Stoermer (2001) proposed that we have now moved out of the Holocene into a new period they labelled the **Anthropocene** based on its distinguishing hallmark, namely human impact on the Earth system. There is no doubt that the dramatic

increase in population and in our use of resources have had major consequences for the Earth's environment in recent centuries (Oldfield, 2005; Chapters 8–10; Steffen et al., 2011). But will human impacts of the kind shown in [Figure 7.1](#) leave a record that is clearly detectable to a geologist living millions of years in the future? For this to be the case, these would need to create Anthropocene strata distinct from the underlying and overlying layers of rock (Zalasiewicz et al., 2011). This will almost certainly be true in places such as cities, where concrete and tarmac will form a characteristic litho-stratigraphic unit, and in some river valleys, where large dams are creating new fluvial beds behind them. If – as seems likely – rising greenhouse gas concentrations lead to significant warming of the climate, then this may eventually cause some of the world's land-based ice sheets, such as Greenland, to melt away. The result would be a sea level rise of 5 m or more, sufficient for deposition of transgressive marine sediments that would be recognisable world-wide. On the other hand, it seems premature to accept a new geological epoch based on something that has not yet happened!

The 'classic' means of defining geological epoch is by using fossils contained in sedimentary rocks, and this approach may be of help in deciding if the Anthropocene should be separated from the Holocene. No new species have appeared in recent centuries, but many have decreased in abundance or changed their distribution as a result of alien introductions. Around 500 species went extinct during the twentieth-century (McNeill, 2000), although most of them will not be preserved in the fossil record. Should this rate of species loss continue through the twenty-first century and beyond, then a future palaeontologist is likely to look back at an Anthropocene Mass Extinction event

comparable to that at the end of the Cretaceous when dinosaurs died out. However, if we were to look for one distinguishing feature of the human epoch that occurs world-wide and is likely to be preserved in the geological record, then it is probably plastics, even if they end up being broken into tiny fragments. Perhaps this new geological epoch should be called the Plasticene, not the Anthropocene!

Figure 7.1 Increasing rates of global human impact since the beginning of the Industrial Revolution (adapted from Steffen et al., 2011; Gehrels and Woodworth, 2013).



It is not easy to agree a date for when the Anthropocene began, if indeed it has done so. Paul Crutzen proposed using the start of the Industrial Era around 1800 AD, when our use of fossil fuels became globally significant. For others, a sharp augmentation in global human impacts is more easily recognizable during the middle of the twentieth century (Figure 7.1). At the other end of the timescale, researchers such as Bill Ruddiman have argued that, as judged from greenhouse gas emissions, the anthropogenic era

began 5000 years ago (see [Technical Box VIII](#)). Still further back in time, the mass extinction of large mammals around the end of the Pleistocene would probably not have occurred without human involvement (see Chapter 3). Hindsight is a great thing, and current geological trends of human causation will doubtless be much clearer in the future than they are now...

One other feature of the time period covered in this chapter that needs mention is the sources of information available, specifically the wide availability of sources other than those provided by proxy methods. No one would think to reconstruct modern European history entirely from archaeological finds, although it is worth remembering that in some parts of the world, written records go back no further than the early twentieth century AD. It was formerly assumed that the same also applied to environmental history. That is, palaeoecological data were thought to be superfluous for recent centuries because environmental histories based on documentary evidence are available instead. In consequence, palaeo-scientists tended to ignore the uppermost layers of their peat or lake sediment cores and instead concentrated their efforts on events further back in time. Work instigated since 1970, most notably by Frank Oldfield (1977, 1983, 2005), has shown that even for historical times, proxy sources can be crucially important. In particular, many contemporary environmental problems such as ‘acid rain’ benefit from an historical perspective but lack adequate documented records. Proxy methods can be applied to environmental changes over decades as well as over millennia, although it has presented some new challenges, notably in dating techniques. Of course, it would be foolish to ignore archival and similar sources of recent environmental history (Sheail, 1980; Hooke and Kain, 1982; Worster, 1993). It is when they are used in combination that historical, archaeological and palaeoenvironmental records provide the fullest insights into how and why people have transformed the natural world around them (Butlin and Roberts, 1995). This chapter does not rely solely on proxy data, therefore, but uses them in tandem with other types of information.

Climatic changes in historical times

It is commonplace to say that no two years have the same weather. Furthermore, years of ‘exceptional’ weather can run together to form more prolonged periods of ‘abnormal’ conditions, such as the prolonged droughts which have affected Africa’s Sahel zone since the late 1960s. But until recently our expectation has

been that these deviations are random events so that over decades and centuries, they will cancel each other out to give an essentially stable climate. Testing the idea of underlying climatic stability is not without its difficulties. Many proxy climatic indicators such as pollen-based vegetation changes or cycles of stream erosion are amenable to non-climatic interpretation as we move forward into historical times.

To help disentangle these different factors, a longer-term perspective may be of value. Has climatic stability been the norm during the Holocene over periods of decades to centuries? The answer must be 'no'. Secular climatic cycles of varying duration up to 1500 years long are recorded as recurrence surfaces within peat bog profiles and in sediment iceberg-rafted from Greenland into the North Atlantic (Bond et al., 1997), via salinity fluctuations in non-outlet lakes (Fritz et al., 1991; Laird et al., 1996), in cores taken through ice caps (Meese et al., 1994; Thompson, 2000) and by many other palaeoclimatic indicators. Some of these climate archives, such as variations in tree-ring widths, can be resolved to individual years (see [Technical Box X](#)). This permits investigation of the potential causes of short-term changes to the climate such as [ENSO](#) events (Cobb et al., 2003). Understanding the climatic changes of the last millennium also has important implications for the thorny but vital question of how global climate is responding to the recent human-induced rise in the level of greenhouse gases in the atmosphere.

TECHNICAL BOX

Technical Box X: Climate from tree rings

Tree ring growth is influenced by environmental conditions, especially near the edges of a species' geographic distribution. Favourable environmental conditions lead to the formation of a broad growth ring, while adverse conditions have the opposite effect. The result is a unique sequence of wide and narrow rings not unlike the bar-codes used for in-store price tags ([Plate 7.1](#)). The main environmental factors which cause ring-width variations are climatic ones, but the critical stress factor will not be the same everywhere. In the Alps, a narrow tree-ring is usually the product of a cold summer while in the American Southwest low rainfall is more likely to be responsible for stunted growth. Tree-rings therefore provide information not only about age (Stokes and Smiley, 1996; see Chapter 2) but also about past climate, via

measurements of ring widths and properties such as wood density and chemical isotope composition, the study of which is known as dendro-climatology (Schweingruber, 1988; Fritz, 1991).

Tree-ring records offer a number of advantages for climate reconstruction, including wide geographic availability in temperate or sub-tropical regions of the world, annual to seasonal resolution, and in-built, high-precision dating (Briffa, 2000). Dendro-climatology places great emphasis on replication (Cook and Kairiukstis, 1990), with at least 10 trees from each species being sampled at a site. Tree-ring science is more challenging in the humid tropics, because a clear growth season is often lacking, and trees such as the baobab (*Adonsonia*) do not possess easily identifiable rings.

Plate 7.1 Year-to-year variations in tree-ring widths, shown in the graph, reflect growing conditions and hence past climate. Reproduced with permission of Fritz Schweingruber and Inst. WSL, Birmensdorf, Switzerland.



Identifying year-to-year (high-frequency) variations in tree ring records is relatively straightforward, but longer (decadal to century scale) climate signals require de-trending of the data to remove the effects of tree growth on ring width, which become narrower

as the tree gets older and bigger. The problem is that de-trending also removes any low-frequency climate signal, in what Ed Cook and colleagues (1995) have called ‘the segment length curse’. Methods such as Regional Curve Standardization (or RCS) have been developed in an attempt to minimize this problem (Briffa et al., 2004), but it remains the case that tree rings are better suited to identifying short-term (year-to-year) variations in weather than they are for long-term (e.g. centennial) trends in climate.

Climate history and global warming

Systematic weather records enable us to reconstruct temperatures over large areas of land and oceans back to about 1850 (Hulme, 1994; Jones et al., 1999). They show an average warming at the Earth’s surface of around 1°C from the mid-nineteenth up to the early twenty-first century (Trenberth et al., 2007), with warming being greatest in arctic regions such as Alaska and northwest Canada. Most of that temperature rise occurred in two steps: the first between 1910 and 1940 and the second since 1980. Almost all of the warmest years on our planet since 1850 have occurred after 1995. But is this warming unprecedented if we take a longer-term perspective? The last great global warming took place at the start of the Holocene, 11 700 years ago, ultimately linked to changes in the Earth’s orbit around the sun (see Chapter 3). On the other hand, external – or ‘boundary’ – conditions of the global climate that are similar to today’s have only been in place during the late Holocene. In consequence, the last millennium is a particularly useful time period if we want to establish the natural range over which climate has varied and whether we have now moved outside this ‘envelope’ (McCarroll, 2010). A similar argument can be applied to the cause(s) of current climate warming. Measurements taken at the top of Mauna Loa in Hawaii since 1958 have shown a marked rise in CO₂ levels in the atmosphere, and greenhouse gases preserved in ice cores place start of the current increase at around AD 1800 (Figure 7.1). On the other hand, there are other naturally occurring causes of climate change over these timescales, including injections of aerosols into the atmosphere from explosive volcanic eruptions and variations in solar activity linked to sunspot cycles. How far did these factors cause the climate to vary prior to the increase in greenhouse gas concentrations two centuries ago?

Evidence for climatic change over past centuries comes from numerous different sources and places. In Europe and East Asia, historical documents provide information about the weather back to medieval times; for example, the timing of cherry blossom in Japan and Korea has been recorded every spring for more than a thousand years! Starting with Gordon Manley’s (1974) compilation of temperature data for central England, researchers have used documentary

sources to reconstruct the pattern of weather across Europe for the last four centuries (Luterbacher et al., 2004; Brázdil et al., 2005). Some of the most important proxy climate data come from sites near to ecological or geological thresholds, such as the upper tree line in mountains, so that regions like the Alps provide rich pickings for climatic reconstruction. High-latitude ecosystems, such as Arctic lakes, have been similarly sensitive to recent temperature changes (Smol et al., 2005), for example, in the duration of their summer ice cover. In addition to biotic indicators such as tree rings, many non-biotic ones such as mountain glaciers have been climate sensitive, although responding to decadal rather than annual variations in weather conditions. Glaciers leave behind a stratigraphic record of their former extent (see Chapter 2), with moraines marking previous glacier advances down valley (see [Plate 7.2](#)). Dating by dendrochronology and early landscape paintings shows that in the Alps, the most recent retreat to modern glacier positions began in the mid-nineteenth century (Rothlisberger and Schneebeli, 1979). Hans Oerlemanns (2005) compiled 169 records of glacier length changes from different parts of the world and calibrated them against mean temperature. His synthesis ([Figure 7.2](#)) shows longer glaciers and – by inference – cooler temperatures prior to around 1900 AD, followed by a two-step pattern of glacier shrinkage that mirrors the pattern of temperature rise recorded using thermometers.

The current phase of retreating glaciers marks the end of a prolonged period of relatively adverse climate in northern and central Europe – the so-called [Little Ice Age \(LIA\)](#) (Grove, 2004). In Europe, the main part of the LIA began around AD 1590 and ended c.AD 1850, although the initial downturn in temperature had begun two or three centuries earlier in late medieval times. On the other hand, it would be wrong to imagine that the whole of the intervening period was cold and damp. Nor should it be thought that the climatic deterioration was anything like so severe as during the Last Ice Age proper, when the drop in mean temperature in some regions was more than 10°C. Even so, the shift in the mean climatic conditions during the centuries of the LIA was sufficient to have measurable environmental consequences. Major rivers like the Thames froze over during some winters ([Plate 7.3](#)), and the effects of climatic deterioration were especially significant in marginal environments such as upland peat bogs. In warmer, drier climatic phases in the eleventh to thirteenth centuries AD, peat bog growth at Bolton Fell Moss in northwest England had been predominantly in the form of dry hummocks. With the subsequent shift to a cooler, wetter climate on the other hand, hummocks gave way to ‘wet lawn’ and eventually to *Sphagnum*-dominated, permanently wet pools (Barber, 1981). Subsequent studies have used microfossils such as testate amoebae (Chapter 2) to reconstruct past water-table depths in peat bogs, a proxy which can then be transformed into records of temperature and rainfall (Charman et al., 2004; Charman, 2010) (see [Figure 6.2](#)).

Plate 7.2 LIA trim lines and moraines (marked), Val d'Hèrens, Swiss Alps. These two glaciers were formerly joined, and, as late as AD 1859, they extended down-valley to beyond the point where the photograph was taken.



The climate was therefore far from constant during the eight centuries before the start of the Industrial era. But how much did temperature vary during that time period relative to the last two centuries? If warming since AD 1850 has been significantly bigger, faster or more widespread than previously, it would give support to the idea that humans are contributing to current climate change. In order to test this, Mike Mann and colleagues (1998, 1999) put together the first systematic large-scale compilation of proxy temperatures for the last millennium. Although it was a [multi-proxy](#) data set, in practice most of their records came from tree rings in mid- to high-latitude regions of the northern hemisphere. These had been calibrated against instrumental measurements for recent decades, which allowed statistical uncertainties to be calculated for their temperature reconstructions over previous centuries. Not surprisingly, these uncertainties got bigger back in time as fewer tree ring and other proxy-climate series became available. Nonetheless, based on their temperature reconstructions, Mann et al. (1999) concluded that it was highly likely (> 90% probability) that temperatures at the end of the twentieth century were warmer than at any time in the last millennium, including periods of warmer climate in medieval times. This was an important conclusion in terms of the global warming debate, and their ice hockey stick-shaped graph was given a prominent airing in the third Intergovernmental Panel on Climate Change (IPCC) report in 2001. The hockey-stick curve has subsequently attracted criticism (e.g. McIntyre and McKittrick, 2005); it was described by the scientific journal *Nature* (441, 1032–3, 2006) as ‘probably the most politicised graph in science’. Subsequent reconstructions have shown larger temperature changes for both medieval times and the LIA than those of Mann et al. (1998, 1999) ([Figure 7.2](#)). Even so, it seems likely that their primary conclusion still holds good, namely, that we are now outside the range of ‘natural’ temperature variability during the last millennium, at least for northern hemisphere land areas (STR Committee of the US National Research Council, 2006).

Figure 7.2 Climatic changes of the last millennium. **A** and **B** Reconstructed northern hemisphere land temperatures; **C** computer-simulated temperatures; **D** principal climate forcings. Adapted from STR Committee of the US National Research Council 2006. Reproduced with permission of the National Academies Press..

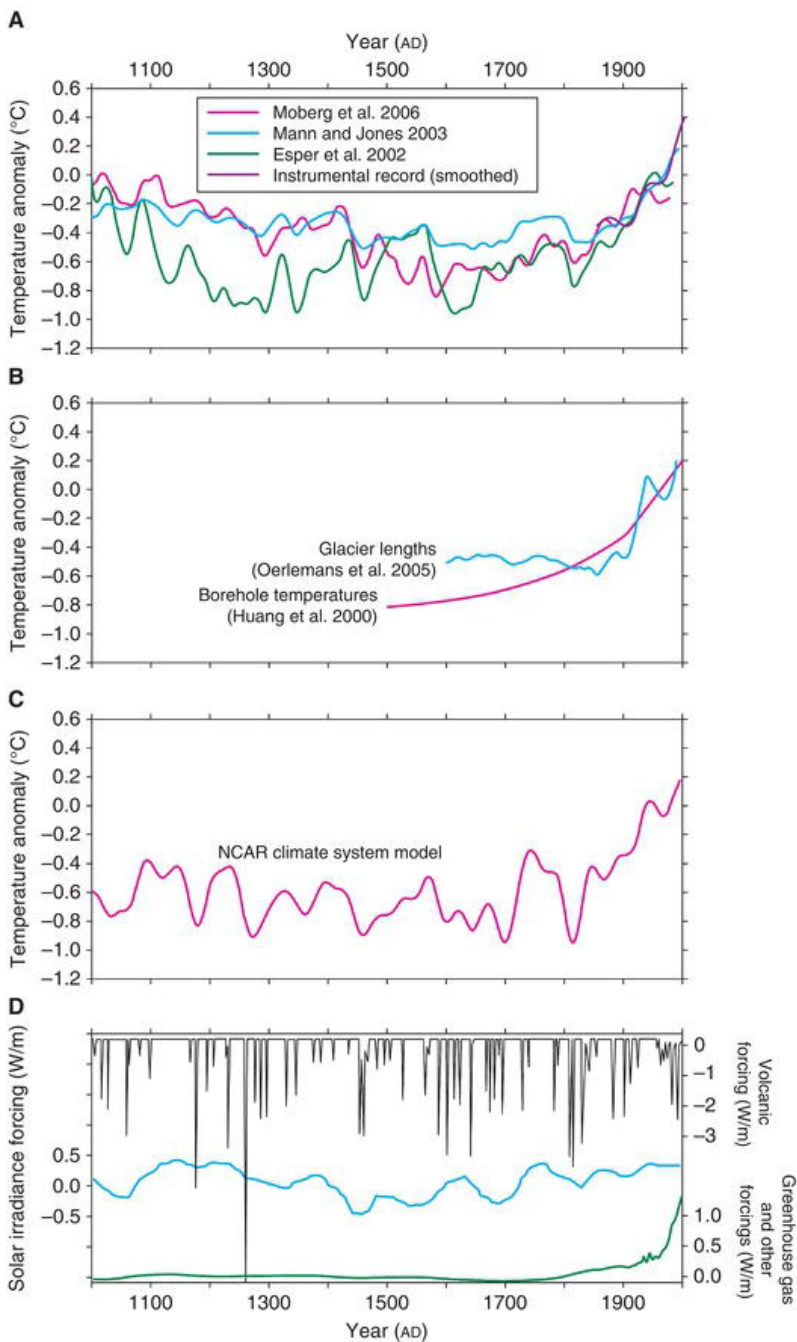


Plate 7.3 The River Thames frozen over in 1895. Earlier during the LIA, frost fairs were held on the solid ice that covered the river surface. Reproduced with permission of the Southwark Local



During the LIA, there were periods such as the ‘Maunder Minimum’ (1645–1715 AD) when sunspot sightings were rare, implying that reduced solar radiation contributed to climatic cooling (Muscheler et al., 2007). Similarly, during the warmer Medieval Climate Anomaly (MCA) around AD 1100 (Bradley et al., 2003), there was an extended period of low volcanic activity. This probably meant fewer sulphate aerosols in the stratosphere, in turn allowing more solar radiation to reach the Earth. Comparing reconstructed proxy temperature curves for last millennium with computer model simulations which incorporate these solar changes and volcanic events (Figure 7.2), there is a good matchup until about 1975, but after this time, warming can only be explained if anthropogenic increases in greenhouse gases are included in the simulations (Hegerl et al., 2007). Temperature reconstructions for the last thousand years therefore provide us with independent information about the sensitivity and natural variability of the climate that can be compared with estimates based on climate models (Jansen et al., 2007; Goose et al., 2008; McCarroll, 2010; PAGES 2k Consortium, 2013). This cross-check is vitally important, because it is computer-based climate models that provide the basis for all projections of the Earth’s future climate.

With two million km² of land lying less than 1 m above current sea level, one of the most profound implications of future global warming will be the risk of increased flooding and erosion in many of the world’s coastal zones. In order to assess how far the rate of sea-level rise has accelerated in recent decades, we need to turn to the record of past sea-level change. As with climate change, the documented historical record – in this case from tide gauges – extends back no more than a few centuries. However, proxy methods can extend this timescale

back into the Holocene. Such methods include the study of foraminifera and testate amoebae, tiny creatures which live at the shoreline and have a range of tolerances for being alternately below seawater and exposed to the air (Gehrels et al., 2001; Charman et al., 2010). Analysis of these sea-level proxy indicators in dated salt-marsh sediment makes it possible to reconstruct rates of sea-level rise over the last millennium. Results not surprisingly show strong geographical variations, because some regions like Scotland are still rising due to glacio-isostatic rebound, while others are sinking due to **isostatic** or to tectonic subsidence. Nonetheless, the overall patterns are consistent in showing that current sea-level rise, which accelerated around AD 1900, represents a significant departure from previous trends during the late Holocene (Milne et al., 2009; Gehrels and Woodworth, 2013) (**Figure 7.1**). As with the hockey stick temperature curve, so also with sea level, it seems as if something unusual is currently happening to planet Earth which cannot simply be explained as ‘natural variability’.

Consequences of Medieval and Little Ice Age climate change

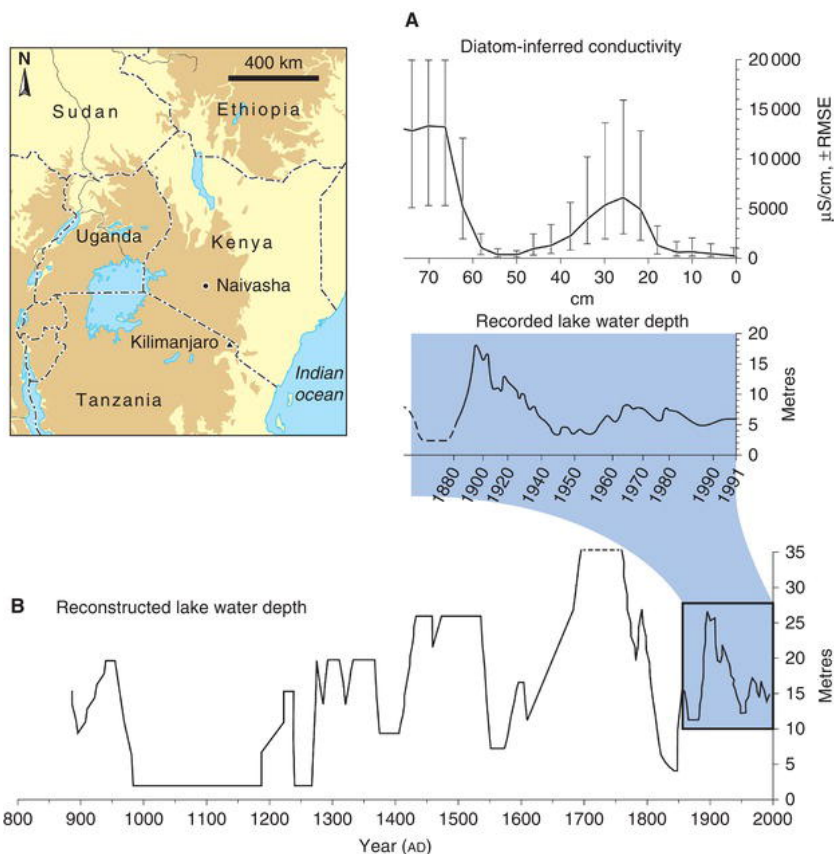
The adverse climatic conditions of the LIA were inimical to domesticated plants as well as to wild ones, and crop failure in Europe’s upland valleys meant that times of feast were replaced by times of famine (Ladurie, 1972; Fagan, 2000). Hill farmers in northern and western Britain no longer found themselves within the agro-climatic limit for cereal cultivation, and many farms and villages in these marginal lands were abandoned between AD 1600 and 1750 (Parry, 1978). Climatic deterioration brought not only poor harvests and pasture but also a higher risk of geomorphic hazards including landslides, avalanches, glacier outbursts and other floods (Grove, 2004). The human consequences of floods, frosts and droughts were often recorded in official reports, diaries and other documents, and these now form valuable source of information about the human consequences of climatic variations in places such as Iceland (Ogilvie, 1992). It can, in fact, become temptingly easy to use climatic fluctuations as a catch-all explanation for socio-economic changes in pre-modern times, when in reality non-environmental factors were at least as important (Redman, 1999; Crumley, 2000; Butzer, 2012). Societies were perfectly capable of ‘bucking’ the climatic trend under appropriate circumstances (Rosen, 2007).

Most evidence for the human consequences of secular climatic variations over the last thousand years comes from temperate areas of the world, most of which have rich historical records or well-studied tree-ring sequences to draw on. But to judge by the experience of the twentieth and twenty-first centuries, drought-prone regions must also have been severely affected by fluctuations in climate, especially in rainfall. Higher rainfall marked parts of the North American Great Plains at the time of the LIA to judge from reductions in inferred lake salinities (Laird et al., 1996), while in California drought phases also seem to have coincided with warmer phases, such as the medieval period, rather than cooling ones (Stine, 1994). Three major droughts between AD 990 and 1300 (Cook et al.,

2004) meant frequent failure of the maize harvest for the Ancestral Puebloan Indians of the American Southwest, and the droughts contributed to the abandonment of settlements such as Mesa Verde (Benson et al., 2007).

Likewise in Africa, Sharon Nicholson (1980) used a range of evidence to conclude that rainfall on the southern (Sahelian) margin of the Sahara was significantly higher at the time of the LIA than before or afterwards. Stratigraphic records of lake levels show that Lake Chad was 4 m above the mid-twentieth-century mean from AD 1570 to 1750, while in the centuries prior to this, both Lake Chad and Lake Bosumtwi in Ghana were at low levels (Talbot and Delibrias, 1977; Verschuren, 2004). This drier phase appears to have been important in causing a retreat of settlement on the inland delta of the River Niger (McIntosh, 1983). Similar hydrological and climatic fluctuations are recorded in the water-level history of East African lakes like Naivasha in Kenya, from which Dirk Verschuren and colleagues (2000) inferred three wetter phases between 1250 and 1800 AD, preceded by an extended dry interval from 1000 to 1150 AD (Figure 7.3). Naivasha also highlights the danger of drawing on short-term experience in developing a sustainable water-use strategy. Settlement by British colonial farmers at the turn of the twentieth century happened to coincide with the wettest climatic period of the last two centuries and a period of high lake levels (Verschuren et al., 1999). These unusually favourable conditions were out of balance with longer-term trends in water-resource availability, and subsequent water usage for cropland has almost certainly become unsustainable without causing damage to ecosystems such as nearby Lake Nakuru, a UNESCO World Heritage site. In lakes like these which lack a surface outlet, water levels rise (and their salt content decreases) at times of high rainfall, and lake levels fall (and their salinity increases) during periods of drought. This was the case, for instance, in Oloidien Bay, which was alternately connected to and isolated from the main lake Naivasha by fluctuating water levels, so causing its salinity to fluctuate dramatically (Figure 7.3A).

Figure 7.3 Lake-level and salinity fluctuations in lake Naivasha, Kenya; A historically recorded water levels and diatom-inferred salinity from a short sediment core in Oloidien Bay (Verschuren et al., 1999; reproduced with permission of the Royal Swedish Academy of Sciences.); B reconstructed water levels for the main lake during the last 1200 years (modified from Verschuren et al., 2000; reproduced with permission of Nature Publishing Group).



Equally dramatic evidence comes from Africa south of the equator. Sediment cores taken in Lake Malawi have shown that this, one of the largest and deepest of all the world's lakes, ceased overflowing and its water level fell by 100 m or more (Owen et al., 1990). This great regression occurred before the start of documented historical data in the mid-nineteenth century but is recorded in local oral traditions. By analysing deepwater cores, Tom Johnson et al. (2001) showed that the last major drop in lake level probably occurred at the end of the eighteenth century. This may have coincided with a period of severe ENSO droughts during the 1790s across much of the tropics, including repeated failure of the Indian monsoon (Grove, 1998). When calibrated climatically, the fall in Malawi's water level represents rainfall only 50–70% of modern values and would have created a drought whose severity and duration make recent climatic events in Africa pale into comparison.

It is clear, then, that recent centuries have been every bit as turbulent climatically in the tropics and sub-tropics as they have been in Europe, with its well-documented LIA. It is equally clear that there has been no simple relationship between variations in temperature and rainfall at a global scale. In lower-latitude regions that receive summer, monsoonal-type rainfall, there

appears to be a match – at least at the timescale of centuries – between periods of warmth and increasing frequency or intensity of drought. This linkage is also given support from analysis of ice cores on tropical mountains, such as Kilimanjaro, where dust flux – and, by inference, dry conditions on the plains below – increased during the warmer period from 900 to 1300 AD (Thompson et al., 2002). On the other hand, in some mid-latitude regions which receive precipitation mainly from winter-season westerly depressions, the medieval period was relatively moist and the LIA was dry, as seen in annually varved sediments from Nar lake in central Turkey (Jones et al., 2006; England et al., 2008). Here the cold, dry winters of the LIA had adverse consequences for grain harvests and sheep herds, which in turn contributed to the gradual decline of the Ottoman empire after AD 1590 (White, 2011).

Expansion at the periphery

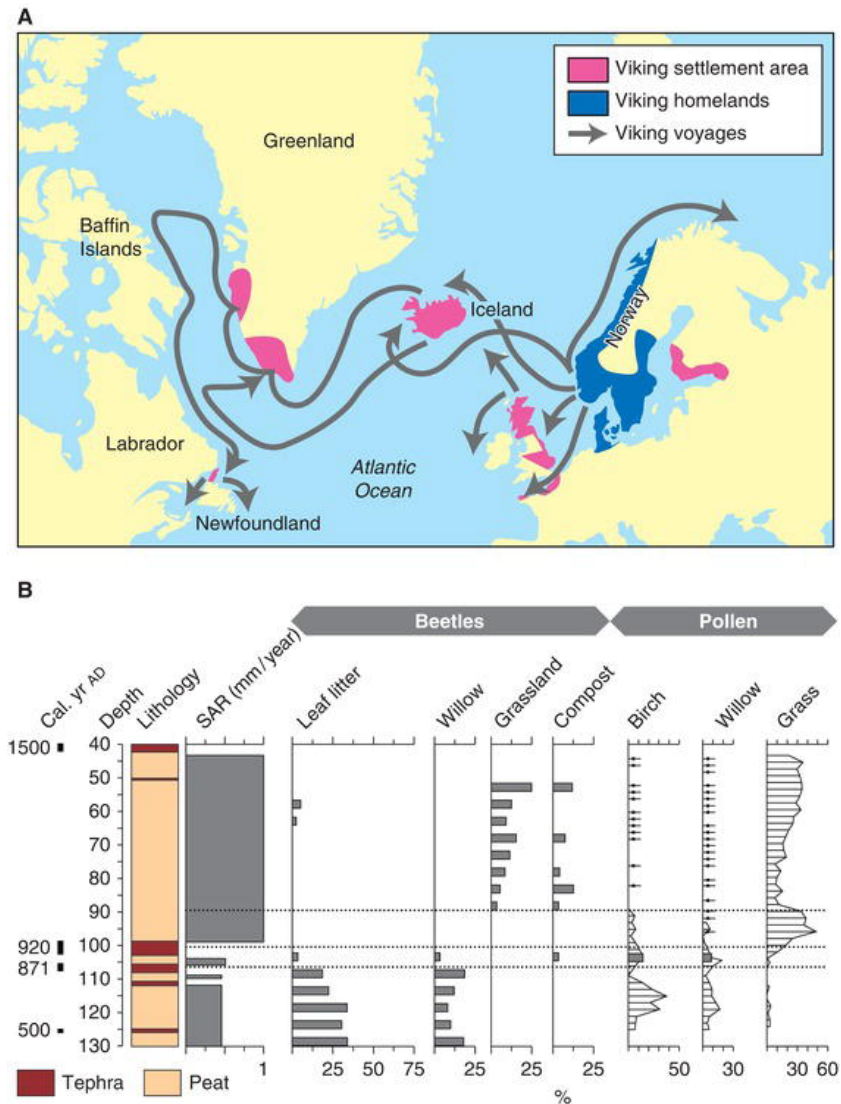
If climatic variations were far from insignificant during the last millennium, what about changes to the natural world on the ground brought about by human agency? They include the continued geographical expansion by human populations into previously uninhabited ‘virgin’ landscapes. The latter part of the Holocene witnessed the last great colonizations of the globe, including the first human settlement of previously uninhabited oceanic islands such as Iceland, New Zealand and Madagascar.

Conquest of the Northlands

The Arctic is one of the most hostile environments for human occupation, and it required specialised cultural adaptation for it to be opened up successfully. After the North American and Scandinavian ice sheets melted away in the early Holocene, tundra developed north of the boreal forest zone around the seasonally frozen Arctic sea. Hunter-fisher-gatherer sites of Pleistocene Age are known from Alaska and Northern Yukon – after all, this was the gateway into the Americas from northeast Asia (see Chapter 3) – but the Arctic north was subsequently bypassed in mainstream cultural developments. In the second half of the Holocene, however, an Arctic maritime culture and technology developed that was specially adapted to the harsh and strongly seasonal periglacial environment (Wohlforth, 2004). Archaeological groups such as the Dorset and Thule cultures were the ancestors of the modern Inuit (Eskimo), and they survived with the same toolkit of snow block houses (igloos), stone lamps, kayaks and harpoon hunting of seal and fish. This human colonisation was aided and abetted by a progressive increase in Arctic sea-ice cover following the mid-Holocene climatic optimum (Funder et al., 2011). On the other side of the pole in northern Eurasia, cultural adaptations were based around reindeer herding, a way of life inherited from Pleistocene hunters, rather than specialised maritime fishing and hunting. The first farming populations penetrated Arctic Scandinavia during Iron Age times, as

testified by the presence of a biomarker for human faeces in lake sediment cores from Norway’s Lofoten islands (D’Anjou et al., 2012) – an unambiguous indicator that people were doing their business there.

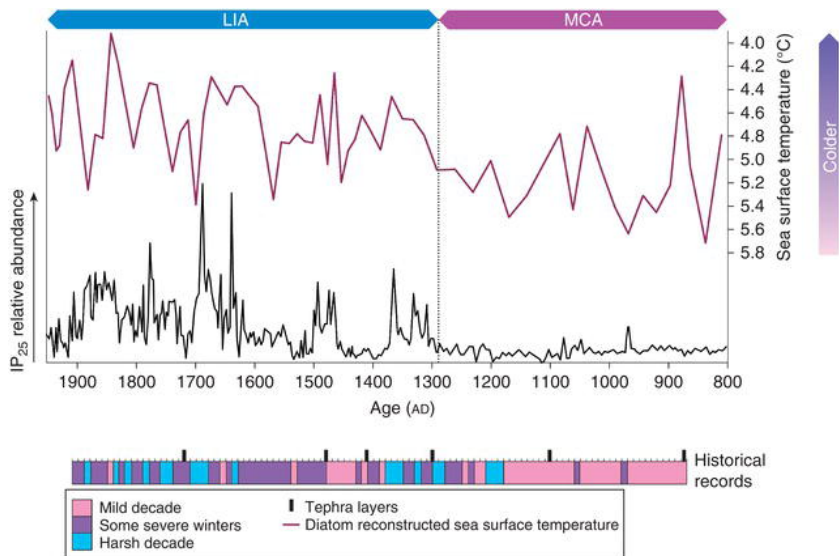
Figure 7.4 A Map showing Norse colonization of the North Atlantic region; B pollen, beetle and sedimentation changes through the Norse *landnám* horizon at Stóra-Mörk, Iceland (data from Vickers et al., 2011). SAR = sediment accumulation rate.



The first major European penetration of the High Arctic took place in the North Atlantic and was carried out by Viking (Norse) settlers who left coastal Scandinavia to reach Iceland, Greenland and eventually Newfoundland in

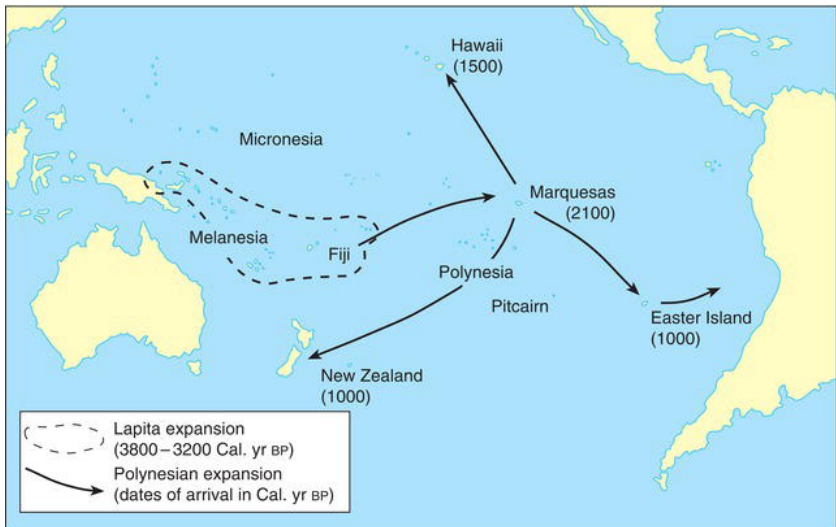
America, between AD 850 and 1000 (Figure 7.4A). Unlike most oceanic islands, those in the North Atlantic possessed no endemic species at the time of the first longship landings. They had been almost completely covered by Pleistocene glaciers which left few, if any, refugia for plants and insects. The biota had, instead, to recolonize the islands at the start of the Holocene, either blown in on the wind or carried as – very cold! – passengers on ice floes drifting across the ocean (Buckland and Dugmore, 1991; Sadler and Skidmore, 1995). Nonetheless by the time of the Norse arrivals, Ári the Wise, writing in the twelfth century AD, was able to look back and record that Iceland was wooded with birch trees ‘from mountain to sea shore’. Iceland soon became home to people and their livestock, along with numerous stowaways such as fieldweeds and grain beetles which had hitched a lift across the ocean. Within 200 to 300 years, there were over 4000 farms, a number comparable to that at the start of the twentieth century (Gerrard, 1991). Human impact on the Icelandic landscape has been dramatic since the initial **landnám** phase, with vegetation clearance and overgrazing by sheep leading to wind and water erosion of the unstable volcanic soils. Tephra layers from Icelandic volcanic eruptions provide convenient marker horizons with which it is possible to calculate how much sediment was eroded, transported and re-deposited for different time intervals. Erosion seems to have started in upland areas, especially on grasslands and heathlands, and only later spread onto the low-lying woodlands (Gerrard, 1985; Dugmore and Buckland, 1991; McGovern et al., 2007). At Stóra-Mörk in southern Iceland, for example, inferred erosion rates doubled after initial human settlement around 900 AD, with pollen, plant macrofossils and beetles remains showing grazed grassland replacing the native birch–willow woodland (Figure 7.4B). The human relationship with nature in this land of ice and fire was not a one-way train, however. For instance, in AD 1783–84 a 30-km-long fissure eruption at Laki led to the loss of an estimated 24% of Iceland’s human and 70% of its animal population, partly because the resulting volcanic gases led to excess fluorine intake (Thordarson and Self, 1993; Edwards et al., 1994; Grattan and Brayshay, 1995). The combined effects of demographic and grazing pressure, an unpredictable climate and fragile soils explain why Iceland’s landscape today is a denuded and barren – if spectacular – one (Plate 6.3).

Figure 7.5 Sea-ice proxy (IP₂₅), sea-surface temperatures and historical climate records around Iceland during the Medieval Climate Anomaly (MCA) and LIA (adapted from Massé et al., 2008).



The inhabitants of the North Atlantic seaboard suffered severely from the climatic deterioration of the LIA, and the incidence of winter sea ice has been documented in Icelandic chronicles for over a thousand years (Ogilvie and Jónsson, 2001), as well as in marine sediment cores (Massé et al., 2008) ([Figure 7.5](#)). Increased sea-salt spray in Greenland ice cores indicates greater North Atlantic storminess after AD 1400 (Dawson et al., 2003) along with the lower temperatures, and this would have hindered communication for the Norse settlements of western Greenland in particular. These communities were always marginal, not only because climatic conditions were poorly suited to agriculture but also because of geographical isolation from their parent cultures in northern Europe. The Greenland Norse seem to have perceived the adverse change in climate during the onset of the LIA as a function of cosmological disorder, and the response on the part of their leaders – who were devout Christians – was to build ever more impressive churches, in order to appease the Almighty. No significant new technological adaptations were made, and the end result was population decline. The western Viking settlement in Greenland was abandoned around 1350 AD, and the eastern settlement followed suit about a century later (McGovern, 1980; Buckland et al., 1996; Barlow et al., 1997; Dugmore et al., 2007, 2012). The Norse were not the only peoples colonizing Greenland, because the Inuit were also moving into these lands from the northwest. Like the Norse, the Thule Inuit cultures had to adapt constantly in order to exploit the available resources; for example, they had to adjust their methods of whale hunting depending on whether the sea ice was close to, or far removed from, the shore (Wohlforth, 2004). There appears to have been little contact between the Norse and the Thule peoples and no cultural exchange, and the former may not even have been aware of the successful Inuit adaptations for use of marine resources. The result was that the Inuit coped more successfully than the Vikings with the climatic deterioration of the LIA, and – unlike the Norse – they survived in the Arctic up to modern times.

Figure 7.6 The late Holocene colonisation of the Pacific.



The Pacific

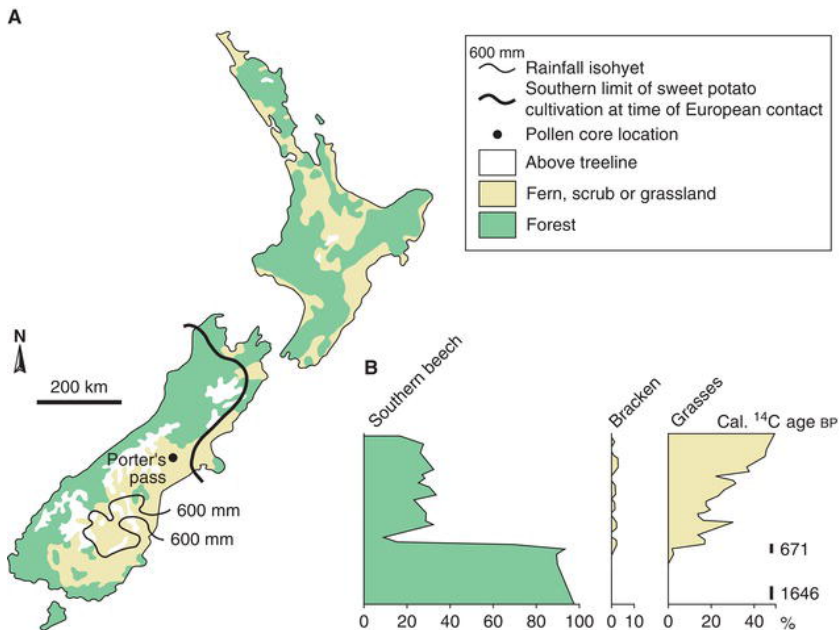
The colonisation of the Pacific was, if anything, an even more remarkable achievement than that of the Arctic Northlands. Prior to 3400 Cal. yr BP, human settlement was restricted to the westernmost islands of Oceania, but in two great surges between then and 800 Cal. yr BP, virtually all of the Pacific was settled by seafaring peoples (Bellwood, 1987; Irwin, 1992; Kirch, 1997, 2000; Kirch and Kahn, 2007; Nunn, 2007). The first surge occurred 3000–2800 Cal. yr BP associated with the expansion of the Lapita culture over Melanesia and western Polynesia (see [Figure 7.6](#)). After a pause of more than a thousand years, a second surge forward took place between 1500 and 800 Cal. yr BP, involving the ancestors of the modern Polynesians. ^{14}C dates on early Polynesian sites and on seeds gnawed by stowaway rats (Wilmshurst et al., 2008) indicate that colonisation pushed three ways from Tonga and Samoa: northwards to Hawaii, southwards to New Zealand and eastwards to the Marquesas and on to Easter Island (Rapa Nui). Genetic analysis of prehistoric human skeletons from Easter Island, Hawaii and the Chatham Islands (near New Zealand) has shown that all share a single mitochondrial DNA lineage (Hagelberg et al., 1994). This confirms that these peoples shared a common origin, despite their enormous geographical spread, and that they passed through a population bottleneck at some stage in their colonisation of remote Oceania. The Polynesian navigators probably even reached South America to bring back the sweet potato (*Ipomoea batatas*) and – if Thor Heyerdahl is to be believed – the idea for the giant stone statues that were erected on Easter Island. There is no doubt that most of these were deliberate, purposeful colonisations and not the result of chance landings by sailors who lost their course. The long voyages must have been organised in advance for they took plants and animals as well as people to new islands. Sharing canoe space were

dogs, chickens and pigs and seeds or tubers of crops including coconut, taro, yam, banana and breadfruit. These formed the main cultivated foodstuffs of island horticulture, which were locally supplemented by wild resources, particularly by fishing. The Pacific islanders knew how to 'harvest' the sea's bounty, and it is probable that the sea acted more like a bridge than as a barrier between the island archipelagos.

Following founding of settlement, the standard portable economic package diversified to fit the resource base available on each island. On large, ecologically diverse islands such as Hawaii, agriculture became highly productive – even involving irrigation. This often came to support a hierarchical society whose tribal chief had as much power over land and life as did any English lord of the manor (Kirch, 2000). By contrast, on smaller and more remote islands, things had a tendency to get lost! The Lapita tradition of pottery making, for instance, died out on most Pacific islands, producing the unusual archaeological sequence of aceramic layers overlying ceramic ones. In many cases, canoes and seafaring skills deteriorated after initial colonisation so that inter-island contact declined and each archipelago became a virtual 'island universe'. Perhaps the ultimate example of cultural reversal occurred on Pitcairn where Polynesian settlement died out completely. Consequently, when the mutineer sailors from HMS *Bounty* landed here in AD 1790, they found coconuts and breadfruit growing, stone platforms and statues, but no people. Small wonder that these became known as the Mystery Islands!

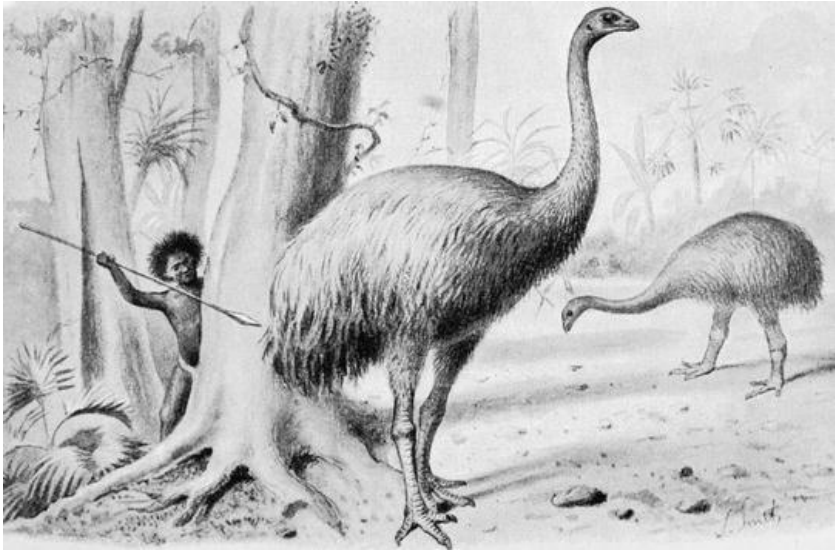
Just as islands provide 'living laboratories' for studying ecological processes, they also provide controlled conditions for assessing human impact on natural ecosystems. Like Iceland, these Pacific Islands were devoid of human population until the late Holocene, although it should be noted that – unlike ecosystems on continental land masses – they lacked indigenous animal populations too. The consequences of the arrival of Polynesian settlers on island ecosystems that had previously been untouched by human hand were far-reaching. New Zealand, colonised by Polynesians about AD 1200 (Newnham et al., 1998), provides an informative example.

Figure 7.7 A Extent of forest in New Zealand c.AD 1850; B pollen record of Polynesian deforestation from Porter's Pass. Adapted from McGlone 1983; reproduced with permission of John Wiley & Sons.



New Zealand has a moist subtropical to cool temperate climate which would seem ideal for forest growth. Yet when Captain James Cook and the other Europeans arrived here, they found that almost half of the land was treeless. A proportion of this lay above the tree line, notably in South Island, but the majority comprised lowland scrub, fern or tussock grassland (see Figure 7.7). However, within the grassland soils could be found abundant charcoal and even wood, suggesting that this had not always been open country. Pollen analyses have since shown that prior to the arrival of Polynesian settlement, only a limited area of central Otago (South Island) with low rainfall failed to support continuous lowland forest cover. Polynesian deforestation was most severe between 800 and 600 Cal. yr BP, but seems to have stabilised before European contact (McGlone, 1983; McGlone et al., 1994). Clearance was effected mostly by burning and does not seem to have been solely or even primarily because of the need for agricultural land. Maori horticulture was concentrated in North Island, where it was warm enough for the main staple crop – sweet potato – to be grown. Yet it was in South Island that deforestation was most severe. This was doubtless partly due to the low rainfall in eastern South Island, which made *Nothofagus* (southern beech) forest especially sensitive to fire and slow to recover after burning (McWethy et al., 2010).

Plate 7.4 Giant flightless moa birds (*Dinornis* sp.) were hunted to extinction by Maori Polynesians after their arrival on New Zealand 800 years ago. Source: Joseph Smit (Public domain), via Wikimedia Commons.



Clearance was almost certainly related primarily to other forms of resource use than farming, notably the desire to encourage the growth of bracken (*Pteridium aquilinum* var. *esculentum*) whose rhizomes were an important source of food. Burning may also have been associated with Polynesian hunting of the flightless moa birds (Dinornithiformes) whose densest populations lay in dry forest and scrub on the eastern side of South Island. The largest of these, *Dinornis giganteus* (Plate 7.4), would have weighed 200 kg or more. ‘In the absence of any other large wild animals’, as Richard Owen noted (1849, p. 270), ‘the whole art and practise of the chase must have concentrated on these unhappy, cursorial birds’. And so it proved. Faced with the loss of their habitat through burning and ruthless human predation, moas and several other types of bird were pushed into extinction only a few centuries after the arrival of the Polynesians (Anderson, 1989; Worthy and Holdaway, 2003). Over-predation of seal and bird populations, destruction of the forest and consequent soil erosion left a degraded landscape and an impoverished economy. Well before European contact, South Island Maori society had gone into eclipse.

Plate 7.5 Easter Island statues. Pollen evidence records how the island was completely deforested following Polynesian arrivals on the island. Photo credit: Hulton-Getty.



An even more poignant example of human impact occurred some 6500 km away on Easter Island (or Rapa Nui). This, the most isolated piece of inhabited land in the world, is renowned for numerous gigantic figures carved in stone and originally set in rows on stone platforms or *ahu* (see [Plate 7.5](#)). They were the work of Polynesian settlers who arrived there at about the same time as their cousins reached New Zealand (Hunt and Lipo, 2006), probably benefiting from strong westerly winds which replace the normally dominant easterlies immediately before a major [ENSO](#) event (Caviedes and Waylen, 1993). Pollen cores from three crater lakes have shown that at the time of the settlers' arrival, Easter Island was largely covered by palm forest (Flenley and King, 1984; Flenley et al., 1991). These pollen records also show that within a few centuries, the palm forests had been removed completely, so that the first European explorers encountered a barren, almost treeless island littered with toppled giant statues from a bygone era.

Paul Bahn and John Flenley (1992) have proposed that this ecological crisis was self-inflicted by a society obsessed with display and rivalry and which accorded monument construction greater importance than conserving the island's finite timber resources. This interpretation has, in turn, been used as a sort of parable to warn the planet as a whole of the dangers of environmental degradation; Jared Diamond (2005), for example, described Rapa Nui as 'the clearest example of a society that destroyed itself by overexploiting its own resources'. However, alternative interpretations are on offer, notably from Terry Hunt (2007). He noted that, as in New Zealand, the main phase of deforestation on Easter Island is now dated to immediately after people first arrived rather than to several centuries later. The ecosystem shock following first human contact may have been due not only to the Polynesian colonists themselves but also to their four-legged accomplices, in this case by stowaway rats who ate the palm nuts,

and so prevented trees from regenerating. According to this explanation, whether deforestation was intentional or accidental, it did not amount to 'ecocide', nor was it the result of over-population. Instead it was caused by the vulnerability to external disturbance of island ecosystems that had evolved in isolation.

Of course, not all Polynesian impacts were as dramatic or injurious as on Easter Island. Many indigenous agricultural systems in the Pacific, such as integrated taro beds, became superbly adapted to local environments (Kirch, 1994; Vitousek et al., 2004; Nunn, 2007). Nevertheless, enough has been said to dispel the European myth of the South Sea islands as an egalitarian paradise in which people and nature lived in perfect harmony. The Polynesian impact on island ecosystems of the Pacific was a severe one.

Ecological imperialism

The 'Industrial Revolution' is usually considered to be the most dramatic economic and cultural shift to have taken place in modern times. With the advent of industrial capitalism, human use of resources became increasingly exploitative, and most daily productive activities became separated from land, climate and the rest of nature. The countryside and wilderness became the escape from routine labour, not part of it. Even agriculture has become industrialised through mechanisation, and land-use activities are now determined more by decisions taken in Brussels or Washington, DC than locally in Galloway or Michigan. For many of the world's peoples and ecosystems, however, the dominant change of recent centuries has been brought about not by industrial capitalism but by European colonial expansion (Arnold, 1996). European culture encountered new and unfamiliar natural environments overseas, and what resulted was unlike either had been before the encounter. Adaptations were often harmonious, as in the vineyards of California, but more than occasionally were not, the dustbowl of the western American prairies providing a classic example of maladaptation of people to – what was for them – new land.

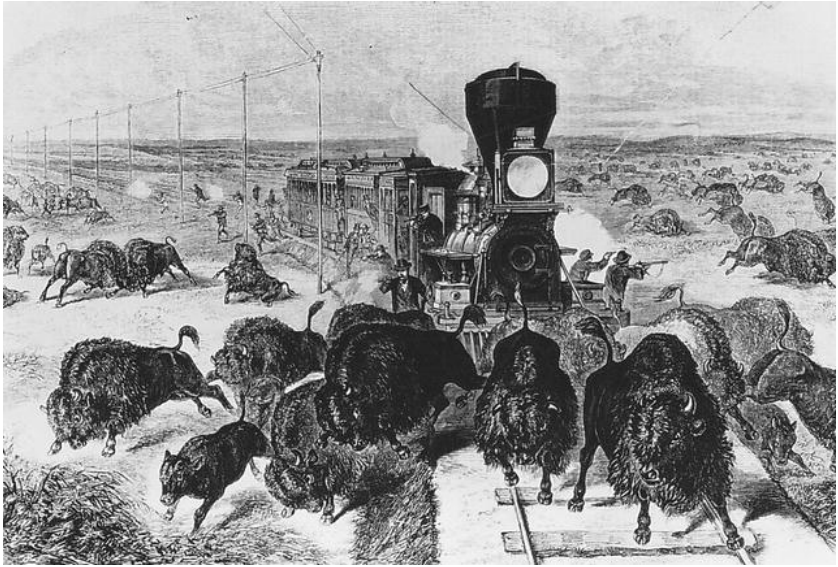
The shift to a modern pattern of human–environment relations began about five centuries ago, when Vasco da Gama rounded the Cape of Good Hope to reach India, and Christopher Columbus set foot on the New World of the Americas. In AD 1450, Eurasia, the Americas, sub-Saharan Africa and Australasia were unknown to each other, but by AD 1550, European explorers had brought all except the last within a system of global contact (Wallerstein, 1974). The explorers returned not only with gold but also with a plant to be 'smoked' and fine bird feathers for ladies' hats. The discovery and subsequent interchange of plants and animal domesticates brought to an end the discrete crop cultures that had existed for so many thousands of years in western and eastern Eurasia and the Americas (see Chapter 5). Indeed, so radical was this transformation that such things as Irish potatoes, Indian palomino ponies, Australian sheep and Canadian wheat are now accepted as native. In fact, all were introduced and adopted following the age of European discovery.

For plant crops, there was two-way traffic between the Old and New Worlds, but the situation was different for animals. The Americas had largely lacked

indigenous domestic animals before AD 1500, and even the wild mammal fauna was species-poor following the wave of megafaunal extinctions at the end of the last glaciation. The ecological potential for intrusive animal species was immense, and the colonisation of the New World was consequently accomplished by four-legged as much as by two-legged Europeans. The omnivorous domestic pig was a particularly effective 'primary colonizer' and was followed by cattle, horses and sheep, some of which escaped to become **feral** species. In the southwest United States, for example, donkeys (*Equus asinus*) used during the nineteenth-century gold rush returned to the wild and are now more numerous than the native pronghorn antelope (*Antilocapra americana*). The feral donkeys are able to outcompete the antelope by virtue of their consumption of a much wider range of graze and browse and are degrading the vegetation as a result.

It has often been the unintended rather than the intended consequences of European contact that have had the most far-reaching ecological effects (Crosby, 1986). Species which had evolved separately over millions of years suddenly came into contact to produce a homogenised, melting pot biota (Buckland et al., 1995). Weedy plants such as plantain (*Plantago*) that are familiar in European pollen diagrams from the Neolithic on expanded their range to disturbed habitats throughout the temperate world. Some native species benefited too; ragweed (*Ambrosia*) was one such beneficiary in North America, and the *Ambrosia* pollen rise is a standard indicator of clearance associated with the advance of settlement across that continent. But just as some creatures profited, others lost out. Before European contact millions of passenger pigeons (*Ectopistes migratorius*) were to be found in the woodlands of eastern North America, and millions of 'buffalo' (*Bison bison*) grazed upon its prairie grasses and herbs (McDonald, 1981). By the end of the nineteenth century, the bison population of the entire continent had been reduced to a few hundred head (see [Plate 7.6](#)), and of the passenger pigeon there was neither sight nor sound, a bird shot out of the sky as its habitat was reduced through forest clearance.

Plate 7.6 The North American bison, previously numbered in their millions, were almost shot to extinction with the coming of the railroad in the nineteenth century. © Bettmann/CORBIS. Reproduced with permission of Corbis Images UK, Ltd.



Extinction rates were highest on island ecosystems, especially on those few – like the Galapagos – which remained uninhabited prior to European contact (Steadman et al., 1991; Turvey, 2009). As Charles Darwin noted on his voyage aboard HMS Beagle, isolation produces unusual, even bizarre, island adaptations including giant flightless birds like the dodo. Once that isolation is broken and competition introduced in the form of pigs or rats carried on board ship, native island biotas are ill adapted for survival, and many species became as dead as the dodo. Richard Grove (1995) has argued that it was observations of species extinction on island ecosystems such as St Helena that prompted the first environmental awareness among Europeans, which in turn led to the development of Western conservation ethos.

The consequences of abrupt external impact are well illustrated by the subtropical Atlantic Islands of the Madeiras. Madeira possessed no indigenous human population when it was discovered and colonised by the Portuguese in the fifteenth century AD, but it did have an almost complete cover of ‘great trees’, including the sharp cedar (*Juniperus oxycedrus*) and the extraordinary and long-lived dragon tree (*Dracaena draco*). Indeed the very name that Madeira was given meant ‘timber’. To the settlers, however, the trees were an impediment and ‘...it was therefore first of all necessary...to set fire to them. By this means they razed a great part of the forest’ (Cadamosto, 1455, quoted in Naval Intelligence Division, 1945, p. 46). Following the conflagration – supposedly seven years long – cattle, sheep and rabbits introduced by the Portuguese prevented woodland regeneration so that today the remnants of the original forests are confined to a few precipitous ravines on the north coast of Madeira’s main island. On the smaller island of Porto Santo, the environmental impact following contact was even more dramatic. Furthermore, it can be traced to a single event, namely, the release onto the island of a female rabbit and her offspring, taken to the island by the father-in-law of Columbus, Bartolomeu Perestrelo. In the absence of local predators, the

rabbits bred prolifically and ‘overspread the land, so that our men could sow nothing that was not destroyed by them’ (quoted in Crosby, 1986, p. 75). Despite attempts by the settlers to kill them off, the rabbits devoured so much of the vegetation and crop cover that severe erosion followed, water sources dried up and the degraded island had to be abandoned for human settlement. The impact of human arrivals extended to many elements of the island ecosystem. For example, its native land snail fauna is – even today – rich and highly endemic, but nonetheless it has been significantly depleted through extinction during the last five centuries (Goodfriend et al., 1994).

Land-use history and soil erosion

Increases in soil erosion have occurred at many times and places during the Holocene, commonly as a result of climatic change (Knox, 2000), but moving towards the present day, human acceleration of the rate of soil loss becomes a dominant influence. In particular, the progressive adoption and intensification of agricultural modes of production have led to widespread conversion of natural woodland or grassland vegetation into farming land (Williams, 2003; Kaplan et al., 2011). Clearance for crop cultivation need not leave the soil completely bare, and some of the original flora may remain, for example, in hedgerows, while the crop itself affords considerable protection from raindrop impact. The invasion of weeds during fallow periods, or of secondary forest after abandonment by tropical shifting agriculture, has a similar protective effect. On grazed grassland, the ground cover may hardly be reduced at all. Nonetheless, agricultural clearance has certainly increased soil erosion losses and consequent sediment flux well above the long-term geological norm (Wilkinson, 2005; Hoffmann et al., 2010).

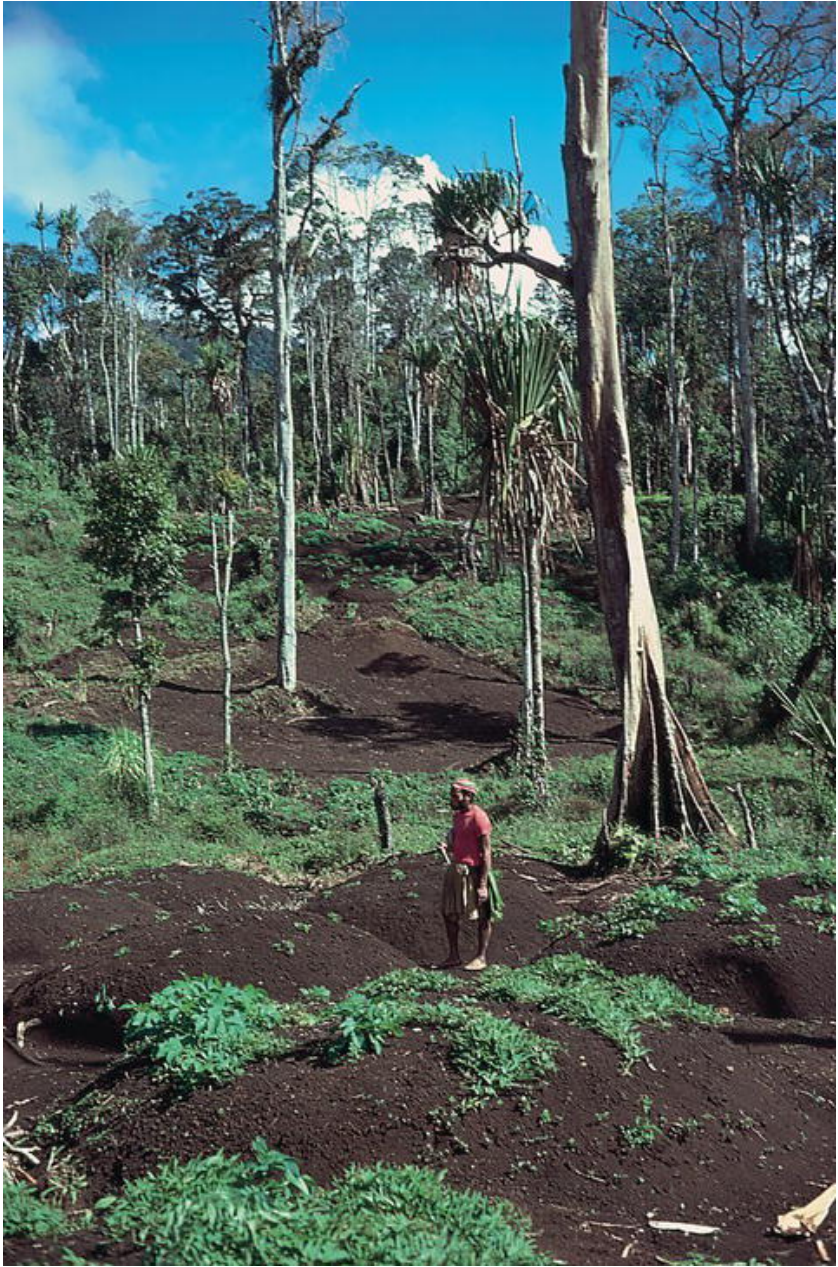
The environmental consequences of soil loss from agro-ecosystems are complex and highly variable (Thornes, 1987; Boardman et al., 1990). Nutrient depletion and other forms of soil degradation can crucially affect conditions for plant growth, changing not only crop yields but also restricting the range of wild plants that could eventually re-occupy the site. One useful indicator of environmental impact is ‘soil life’, or the balance between the rate of soil loss and the creation of new soil by substrate weathering. Soils on alluvial floodplains, for example, may take only 50 years to form and are generally subject to low rates of erosion, making their soil life effectively infinite. In contrast, soils on the ancient land surfaces of Africa have taken millions of years to form and in some cases have lives as short as the rate at which they are being lost, which can be as little as 10 years on soils actively eroding today. Nor is impact restricted to the site of degradation. Eroded soils are washed down slope and downstream to push up the sediment loads carried by rivers and affect their aquatic life.

Human beings are not passive to land degradation, and conservation measures to maintain soil fertility are surely as old as agriculture itself. Traditionally,

tropical shifting cultivators have left tree stumps in their fields and intercropped to reduce the impact of the violent convectional rains on the soil surface (see [Plate 7.7](#)). Manuring, which provides nutrients and helps maintain soil organic matter levels, is another very ancient practice (Wilkinson, 1989). The laborious work of terracing China's loess plateau by peasant farmers has greatly reduced erosion in an environment where natural soil losses were amongst the highest in the world. However, there may be a long-term price to pay for cultivating marginal land such as this. Soil conservation systems need maintenance, and if they break down, it can have environmental consequences worse than if the land had been left alone in the first place.

Over the past 500 years, European expansion and colonialism have brought many important changes in land use and traditional culture–environment relations. In some continents, such as the Americas and Australia, this involved wholesale transfer of European populations with the consequent eclipse or demise of indigenous peoples (Turner and Butzer, 1992; Flannery, 1994, 2001). In other parts of the world, especially in the tropics, European contact left the indigenous peoples intact, but disrupted their established mode of production and threatened their traditional relations with nature. 'You ask me to plough the ground. Shall I take a knife and tear my mother's bosom?' was the reaction of the Amerindian Smohalla to the introduction of Old World agricultural technology.

Plate 7.7 Shifting cultivators in the tropics often leave the larger trees in place; their leaves intercept the erosive rainfall and their roots bind the soil together, helping to minimize soil loss. Reproduced with permission of Timothy Bayliss-Smith.



In order to establish the impact that European cultural contact had on the natural environment, three case histories of the relationship between land-use history and soil erosion in different parts of the world are discussed here. The first example derives from 'Old Europe' and serves as a yardstick for comparison with

nineteenth-century European settlement and clearance in the eastern United States and twentieth-century 'colonial' impact in the highlands of Papua New Guinea. Lake-based studies provide the main source of proxy soil erosion data in these examples. Lake sediments are of particular value because the same cores used to reconstruct sediment influx and catchment erosion rates can also provide evidence of land-use changes via their pollen record (Oldfield and Clark, 1990; Dearing and Jones, 2003; Enters et al., 2008). In considering these examples, it is more important to contrast different trends through time than to compare the specific values computed for past erosion rates, for these are partly determined by local factors such as geology and lake catchment size.

Havgårdssjön is a relatively shallow, kettle-hole lake situated in an area of predominantly arable farming in southern Sweden. Over recent centuries, the main change in land use has been agricultural intensification and in particular the conversion of pasture into arable land. John Dearing calculated past erosion rates by taking a series of 1-m-long sediment cores set in a grid over the lake surface. Dry sediment influx rates and mean catchment erosion losses were then produced for different time periods during the last 350 years (see [Figure 7.8](#)). The main change recorded is a modest increase in the inferred erosion rate during the nineteenth century. Study of historical records has shown that this coincided with an expansion in the area of ploughed land and a sharp increase in the sheep population. Although this agricultural intensification appears to have increased soil losses from the catchment, erosion rates were already relatively high even in the seventeenth century. This is surely related to the fact that clearance of the original woodland cover took place before the modern period, especially in Viking times (Dearing et al., 1987). A longer sediment core from the lake does show a progressive increase in the sediment accumulation rate from 5000 Cal. yr BP onwards as do many other northwest European records (Dearing, 1994). At Holzmaar, a varved lake sequence from western Germany, sediment influx increased in a series of abrupt steps at around 2800, 2050 and 950 Cal. yr BP associated with Iron Age, Roman and medieval woodland clearance, particularly for charcoal and iron production (Zolitschka, 1998). In this small lake catchment, soil losses under pre-modern landuse represented a twenty-fold increase compared with that under early Holocene forest cover. Without doubt, land degradation has occurred in temperate Europe in recent centuries related to population growth and agrarian pressure (Dodgshon, 1988; Higgitt et al., 1991; Lang, 2003; Enters et al., 2008), the same pressure which contributed to the subsequent European expansion overseas. But land clearance and degradation have a long antiquity in this corner of the world, and records such as those from Havgårdssjön and Holzmaar can only be understood by considering changes in prehistoric and early historic periods, as well as in more recent times.

Figure 7.8 Comparative soil erosion rates reconstructed from lake sediments: Sweden, Michigan and Papua New Guinea (data from Dearing et al., 1987; Davis, 1976b; Oldfield et al., 1985).

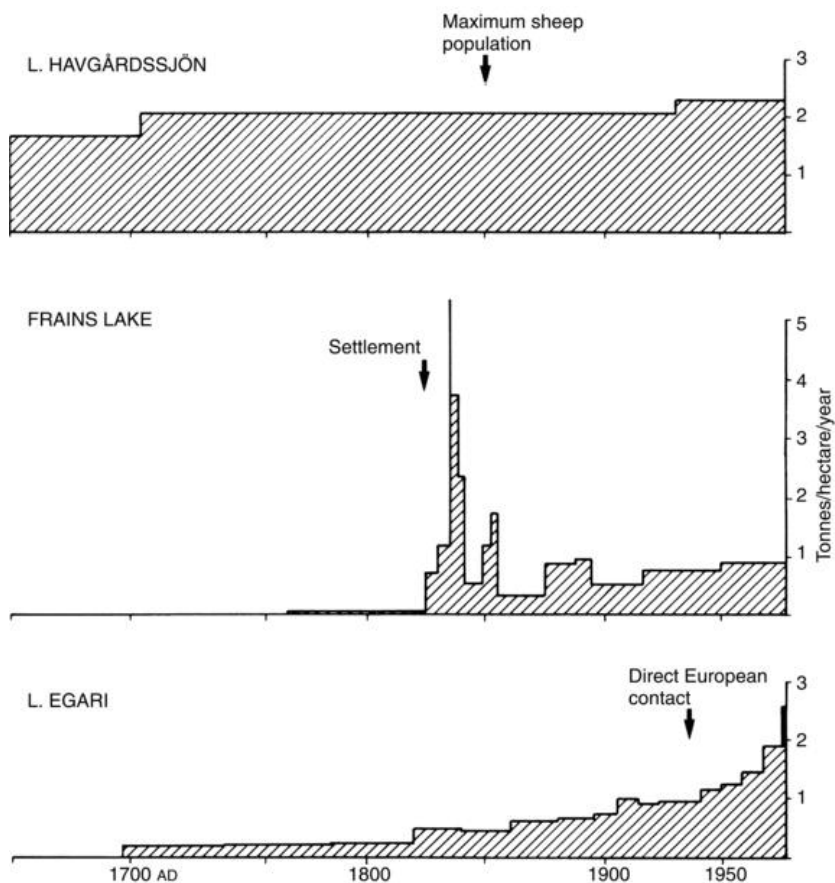


Table 7.1 Comparative statistics for three lakes used to reconstruct catchment erosion rates

	Håvgardssjön	Frains Lake	Lake Egari
Altitude (m above SL)	51	260	1800
Lake area (ha)	55	6.7	8.3
Catchment area (ha) ^a	141	18.4	9.2
z (lake:catchment) ratio	0.39	0.36	0.9
Max. water depth (m)	c.5.5	c.10	11
Core density (per ha)	1	3.3	0.6
Core correlation	m.s.	pollen	m.s., volcanic ashes

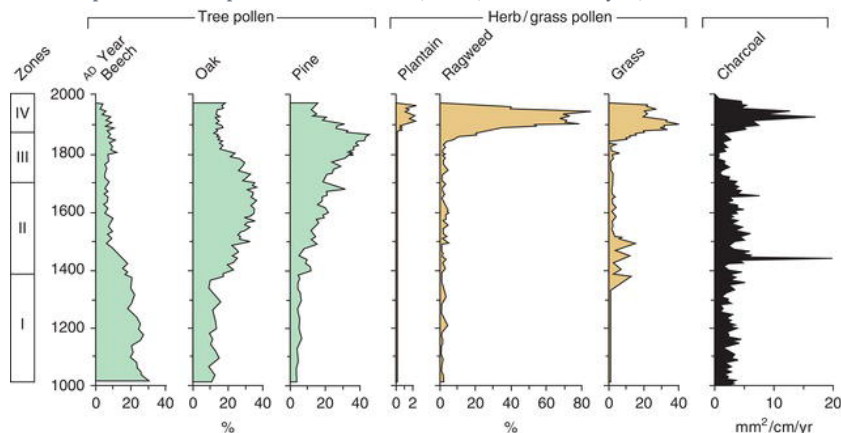
^a Excluding lake
m.s., magnetic susceptibility

Frains Lake, Michigan, is somewhat smaller than Håvgardssjön, but has a similar lake:catchment area (z) ratio (see [Table 7.1](#)). As in southern Sweden, the natural vegetation around Frains Lake would have been mixed temperate woodland, but in Michigan, it was cleared for agriculture much more recently, mainly in the nineteenth century AD. How would the sudden arrival of European settlement affect erosion in the Frains Lake catchment? Margaret Bryan Davis (1976) reconstructed the land-use and erosion history from sediment cores which were cross-correlated by pollen analysis and sediment stratigraphy. The modern settlement was known to have been founded in AD 1830, and the associated forest clearance showed up as a clear landnám phase in pollen diagrams, with a sharp rise in ragweed (*Ambrosia*) pollen. The same horizon was also marked by a change in sedimentation from organic gyttja mud to inorganic clay, reflecting the inwash of minerogenic material following deforestation.

One result of her study was to show that in a small area at the centre of the basin, sediment had accumulated 10 times faster than elsewhere on the lake bed. The fact that no less than 3.7 m of sediment could build up in a localised part of the lake in the 140 years after AD 1830 graphically illustrates the need to use multiple cores in any lake-based reconstruction of total sediment influx rates. Sediment core samples were burnt to produce ash weight from which the influx rate of inorganic material over the last 180 years was calculated (see [Figure 7.8](#)). This showed a dramatic jump in erosion from around 0.1 to over 5 t/ha/year with forest clearance, before stabilising at a new equilibrium level at around 0.9 t/ha/year once farming was established. Longer cores showed that sediment accumulation rates had been stable and low over the remainder of the Holocene, although a slight rise occurred after 2000 Cal. yr BP, possibly related to native American land-use activities. In contrast to Håvgardssjön, the sequence from Frains Lake shows a 10-fold increase in mean sediment yield in recent centuries, and rather than rising gradually, erosion rates rose abruptly upon European impact. Over the mid-latitude portion of eastern North America as a whole, lake sedimentation rates have increased threefold as a result of European vegetation clearance and other impacts (Webb and Webb, 1988). Perhaps the most significant result of the Frains Lake study was in identifying the importance of the

unstable transitional phase between different land uses: in this case forest and farmland. In **ergodically** based studies (e.g. Wolman, 1967), retrodicted erosion rates under different modern land-use types can fail to recognise the importance of instability associated with the transitional deforestation phase.

Figure 7.9 Pollen and charcoal record for the last thousand years from Crawford Lake, Ontario, Canada (adapted from Campbell and McAndrews, 1993; Clark and Royall, 1995).



Similar sequences to that at Frains Lake have been recorded elsewhere in North America (e.g. Plater et al., 2006). In a varved sediment record from Crawford Lake in southern Ontario, the impact of nineteenth-century European settlement on the mixed temperate forest can immediately be seen from pollen studies carried out by ‘Jock’ McAndrews (Figure 7.9). In the uppermost pollen zone IV, there is a dramatic increase in the pollen of ragweed, grasses and plantain as the forests were cut down or burnt, the latter being reflected in a marked increase of charcoal in the lake sediments. Interestingly, this is not the first period of disturbance to the forest that is recorded at Crawford Lake, although it is certainly the largest. An earlier clearance phase in the fourteenth and fifteenth centuries AD (pollen zone II) was linked to native Amerindian settlement and maize cultivation around the lake (Finlayson and Byrne, 1975), which also led to lake eutrophication (Ekdahl et al., 2007). The associated change in forest composition from beech to oak to pine dominance may have been related to the increase in forest fires resulting from burning by Iroquois Indians (Clark, 1995; Clark and Royall, 1995); alternatively – or additionally – it may have been a product of LIA cooling (Campbell and McAndrews, 1993).

In another study that linked environmental changes recorded in sediment cores to LIA climatic conditions, Grace Brush (1994) found a slowdown in sedimentation rates in Chesapeake Bay during the centuries prior to c.AD 1786. The reduction of catchment forest fires and consequent lack of erosional disturbance along the eastern seaboard of the United States may also have been linked to the massive reduction in Amerindian population caused by smallpox and other ‘European’ diseases (Roberts, 1989; Nevle and Bird, 2008). Diseases emptied many New World landscapes of their native populations well in advance of the main arrival of immigrants from across the Atlantic and allowed forests that had

previously been managed to regenerate (Whitney, 1994; Heckenberger et al., 2003). The main wave of forest clearance by European settlers around Chesapeake Bay was marked by a sharp increase in the rate of sediment flux. Frank Oldfield was able to use magnetic ‘signatures’ to separate the two main sources of sediment into this estuary, namely, bedrock eroded along its cliffs and eroded catchment topsoils (Thompson and Oldfield, 1986; Chapter 10). These components could be identified in sediment cores from the central channels of the estuary, which revealed that whereas bedrock formed the main sediment input before AD 1600, this was subsequently replaced by eroded topsoil, which became the main source of sediment by AD 1880.

Our third case-study area is the **highlands of Papua New Guinea**, and it offers both a different cultural context and natural environment. It experiences intense tropical rains and has steeply sloping, erodible soils, which have been protected from rainsplash erosion by dense montane rainforest. These highlands have been thought of as culturally isolated, being bypassed by nineteenth-century colonialism, and only falling within a ‘Western’ orbit during the 1930s. Over the last 50 years, the traditional mode of production had been disrupted by the introduction of a cash economy. Severe soil stripping is evident in many parts of the highlands today, but it is unclear whether this is connected to twentieth-century contact or to a much longer indigenous process of agricultural intensification. Agriculture began in the Papua New Guinea highlands as early as 9000 years ago (see Chapter 5), involving both swamp and dryland cultivation of yams, taro, sugar cane and bananas and husbandry of pigs. Over the course of the Holocene, it gradually intensified to support one of the highest population densities of any tropical swidden cultivators. There was, as Jack Golson (1977) put it, ‘no room left at the top’ by the time of European contact. What impact did intensive indigenous agriculture have on land degradation in this fragile environment?

Table 7.2 Past erosion rates at Yeni Swamp, Papua New Guinea

Source: Gillieson et al. (1986). Reproduced with permission of John Wiley & Sons.

Time span (Cal. yr bp)	Catchment erosion(t/ha/year)	Land use
5700–3750	0.12	Initial forest clearance
3750–1000	0.09	Swamp margin irrigation
1000–500	0.20	Further clearance with dryland horticulture
500–present	0.61	Intensified horticulture with sweet potato, followed by

European impact

Studies based on the sediment stratigraphy of swamp and cave deposits show a progressive increase in soil erosion rates during the later Holocene (see [Table 7.2](#)). Soil losses were lowest under raised-bed wetland horticulture and rose after 500 Cal. yr BP with the introduction of the sweet potato (*I. batatas*). This crop was swiftly adopted on account of its wide environmental tolerance and quick growing time, and it permitted a demographic growth and intensification of small-scale horticulture, just as the potato (*Solanum tuberosum*) did when it was introduced into Ireland. Both of these tubers are South American domesticates, with the sweet potato probably brought to the Philippines by the Spaniards in the sixteenth century AD. The fact that it reached the highlanders of Papua New Guinea soon afterwards showed that their pre-European cultural isolation was by no means complete (Allen and Crittenden, 1987). Intensive garden cultivation may have led to soil nutrient depletion, which would provide an origin for the grasslands found in some parts of the New Guinea highlands.

Although soil degradation did take place in the past, especially after 500 Cal. yr BP, erosion losses before European contact are less than those recorded after clearance in the much more robust environments of northwest Europe and eastern United States. The link between contemporary erosion and cash cropping would appear to be confirmed by lake-based studies of past erosion rates in the Papua New Guinea highlands (Oldfield et al., 1985). At Lake Egari, reconstructed sediment accumulation rates show a 10-fold increase in erosion rates over the last 150 years, with a doubling since the 1930s (see [Figure 7.8](#)). Although there was no new occupation of land by Europeans as there was at Frains Lake, the environmental consequences of European contact have been similar in both cases. It is a remarkable fact that traditional swidden and wetland agriculture operated in the ecologically fragile highlands of Papua New Guinea for over 8000 years, eventually supporting almost a million people, without serious environmental degradation. This situation only changed when indigenous environmental relations were disrupted, firstly with the introduction of a new exotic domesticate – the sweet potato – and secondly with the advent of the twentieth-century cash economy.

Pollution histories

If the hallmark of agricultural impact has been clearance of natural vegetation to make way for field or pasture, the hallmark of industry has surely been pollution. Industrial society has modified the cycling of natural elements to an extent that was never possible under agricultural modes of production. The use of fossil fuels has reduced the dependence of industrial society on biological energy fixation through green plants and has profoundly altered the global carbon cycle as a result. Other elemental cycles which have been altered include those of key micronutrients such as phosphorus and toxins such as mercury.

The term pollution implies disturbance of the natural environment by putting

the wrong substances in the wrong quantities in the wrong places. Although pollution includes artificial elements and compounds, such as radioactive caesium-137 or toxic DDT, it more usually involves naturally occurring ones like sulphur or nitrogen. This makes it difficult to establish whether the present state of an ecosystem has been modified by pollution or not. Lake Erie is much richer in nutrients than Lake Superior, but is this because of sewage and industrial effluent from around Lake Erie's shores or because the geology of the two lake catchments is different anyhow? A second problem that bedevils pollution studies is that pollutants move downstream or downwind from their point of origin to affect other localities. Establishing the source of pollution, and consequently the cause of it, can be a tricky business. Palaeolimnological techniques cannot overcome these obstacles on their own, but they do have some crucial advantages over other approaches (Smol, 2008). They provide longer and more continuous histories of water chemistry and biology than monitored records and smooth out the random variations of individual years. They further enable the stable baseline condition of the ecosystem to be established, if one ever existed.

In the remainder of this chapter, the environmental histories of some industrial-age pollutants and their consequences for biota are described. In particular, it is shown how palaeolimnological techniques have been used to trace the causes of eutrophication and acidification in freshwater ecosystems.

Eutrophication: natural or cultural?

Ecosystems can be nutrient-rich (eutrophic) or nutrient-poor (oligotrophic) and may change naturally from one to the other. In lakes, it has long been thought that eutrophication occurs through the natural ageing process, as the lake is infilled with sediment and the volume of the hypolimnion is reduced (see [Technical Box IV](#)). **Eutrophication** can also be induced by human activities such as the application and subsequent runoff of chemical fertilizers to farmland or by the discharge of sewage or industrial waste into streams and rivers. In all of these cases, so-called cultural eutrophication is an unintended, indirect consequence of human land-use, sanitation or industrial activities. Nutrients, particularly nitrate and phosphate, are washed away by runoff and carried downstream into rivers and lakes. Most organisms are subject to Liebig's Law of the Minimum, by which productivity will be restricted by the availability of the substance that, in relation to the needs of each organism, is least abundant in the environment (Hutchinson, 1973). In aquatic ecosystems, nitrogen and phosphorus are usually the limiting substances which hold in check greater primary productivity in the form of algal growth.

At first sight, it might appear that eutrophication should be a beneficial process, because it encourages greater biological productivity. Farmers, after all, apply nitrogen and phosphorus to their fields precisely in order to stimulate more prolific plant growth. Unfortunately, the downstream consequences are not always so desirable. Firstly, many organisms are adapted to oligotrophic conditions; the natural habitat of trout, for example, is in clear, if relatively infertile, upland streams and lakes, rather than in murky, eutrophic lowland

waters. Secondly, blue-green algae, which are rather inedible by zooplankton, increase more rapidly with eutrophication than other algal forms. Thirdly, eutrophication can go too far, to produce a hyper-eutrophic environment. Primary productivity may increase to the point where dead algae use up most of the oxygen as they decompose while sinking to the lake bed. Lake waters become de-oxygenated, and higher organisms such as fish, instead of thriving on an abundant food supply, find themselves restricted to the upper few metres of the lake where they are not starved of oxygen. In some cases, such as many of the Norfolk Broads in eastern England, hyper-eutrophic conditions have eliminated stabilising aquatic macrophytes such as *Chara* and *Najas*, and their waters have become smelly and opaque (Moss, 1998).

Plate 7.8 Freeze core taken in annually laminated (varved) sediments from Baldeggersee in Switzerland's Alpine foreland. The change from light to dark sediments dates to AD 1885 and marks the onset of anoxia in the lake's bottom waters brought about by cultural eutrophication (after Lotter et al., 1997). Reproduced with permission of Andy Lotter.

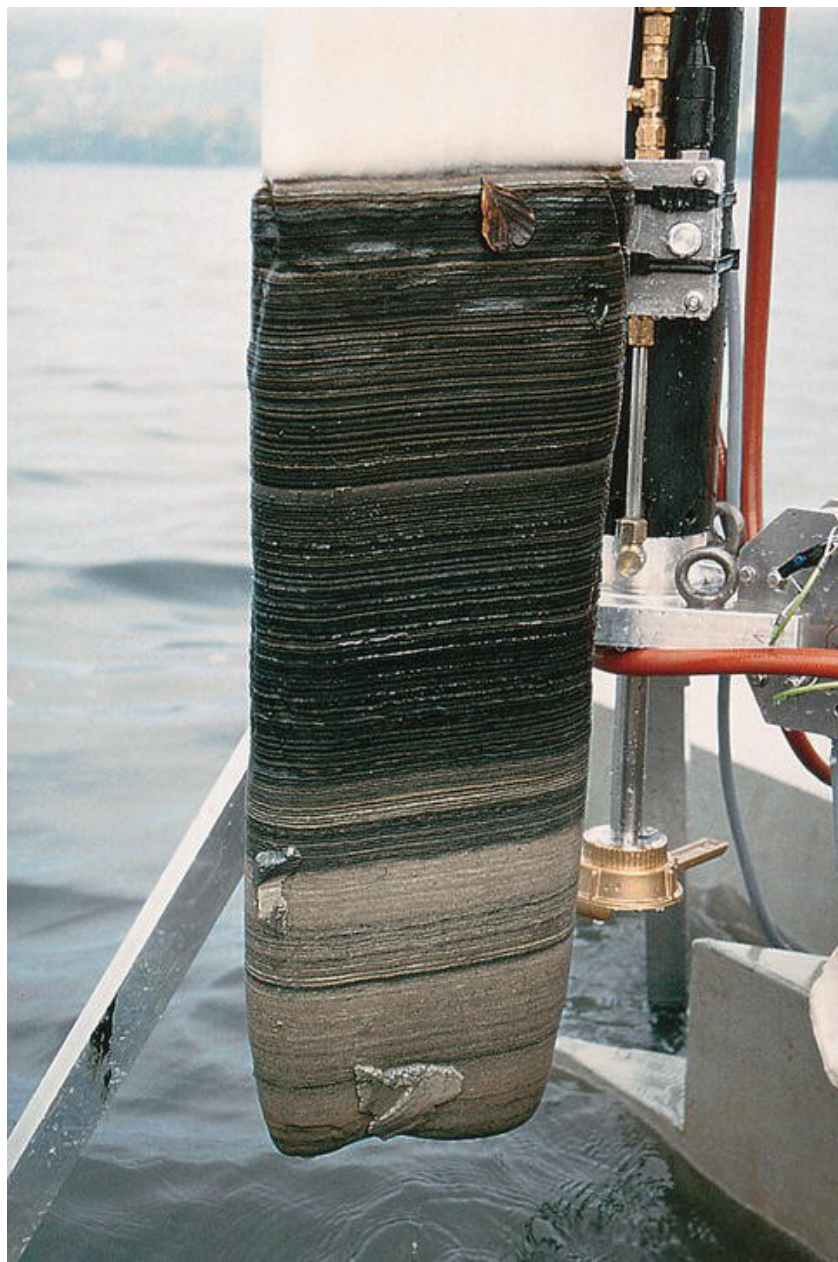
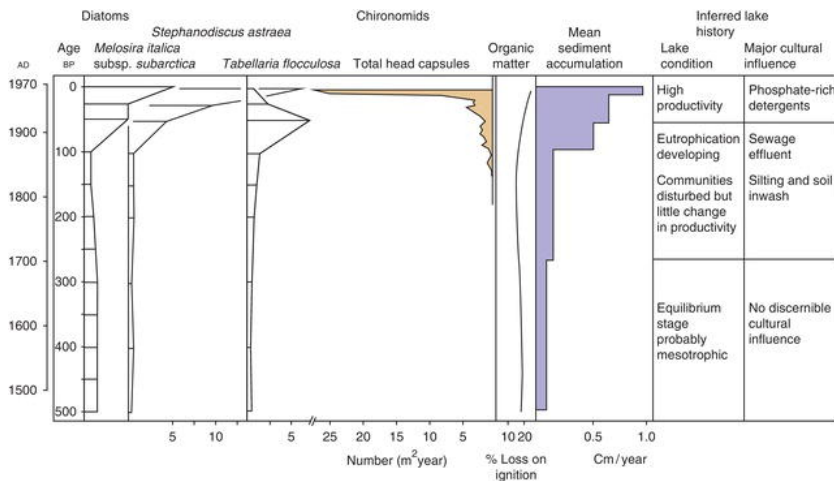


Figure 7.10 Lake sediment record of eutrophication, Lough Neagh, Ireland (adapted from Carter, 1977; Battarbee, 1978).



The sediment profile of a presently eutrophic lake can be used to establish if this has been the long-term, stable state of the ecosystem, or if it has eutrophied over time. In studies of recent lake history (<500 years), only the upper metre of sediment normally needs to be cored, making fieldwork a relatively easy task (see Plate 7.8). One simple measure of former lake productivity is the organic matter content of the sediment profile. Under eutrophic conditions, more organic matter is produced within the lake than is oxidised as it descends to the mud–water interface. On the other hand, an up core increase in organic matter could be caused by changes other than eutrophication – by erosion of peat from the lake catchment, for example. Organic matter changes alone cannot be conclusive. A second, and perhaps more obvious, measure is the chemical composition of the profile, notably its phosphorus and nitrogen content. This has been successfully analysed in cores from a number of lakes such as Lake Washington near Seattle (Edmondson, 1974). However, phosphorus can be easily exchanged between the topmost sediments and the lake water, and this mobility makes phosphorus a somewhat less useful trophic indicator than it might be otherwise (Engstrom and Wright, 1984). A third measure comes from biological indicators of trophic status. Most aquatic organisms are sensitive to nutrient availability and some, although by no means all, are subsequently preserved in lake bottom muds. Amongst the most useful indicators are chironomid larvae (Langdon et al., 2006), cladoceran faunas (Frey, 1986; Korhola and Rautio, 2001) and diatoms (Smol and Stoermer, 2010). With eutrophication, planktonic organisms typically become more abundant and benthic (bottom-dwelling) types become less common, as the hypolimnion becomes deoxygenated. As with most palaeolimnological studies, a combination of different physical, chemical and biological analyses is more powerful than any one technique on its own.

The largest lake in Ireland is Lough Neagh, and today it is markedly eutrophic in character. Its present condition is thought to be linked to the input of pollution by domestic and industrial effluent, although the lake's catchment has a long history of agricultural land use which may also have contributed to nutrient enrichment. In order to establish the origins of Lough Neagh's eutrophic state, a

number of palaeolimnological investigations have been undertaken (e.g. Battarbee, 1978). Organic matter does increase somewhat towards the top of sediment cores obtained from the lake (see [Figure 7.10](#)), but more convincing evidence of trophic change comes from biological indicators, notably diatoms and chironomids. Both organisms show significant changes in species composition in the top metre of the lake sediment profiles, indicating that the present-day lake biota is not a long-term stable one. Rick Battarbee went a step further and reconstructed the palaeo-productivity of Lough Neagh through absolute abundances of core diatoms as well as via percentage counts. Diatoms, of course, form part of the total algal population and their total numbers would be expected to increase with eutrophication. Good dating control is essential in any study of 'absolute' microfossil numbers, and in this case, it was provided through the radioisotopes ^{14}C , ^{210}Pb and ^{137}Cs . This chronology showed that there had been a marked acceleration in the rate of lake sedimentation after about AD 1700.

The results from Lough Neagh show a dramatic increase in total diatom abundance during the late nineteenth and twentieth centuries, with values 10 times higher between 1960 and 1970 than they were during the stable baseline state before AD 1700. There are a number of factors which could prevent these diatom influx figures from being a reliable measure of former lake productivity. These include problems of diatom frustule dissolution, redistribution within the lake and loss down the outflow river, as well as the fact that many other organisms contribute to primary productivity that are not preserved in the lake sediment record. Fortunately, the diatom data are given strong support by a comparable upcore increase in the number of chironomid head capsules (Carter, 1977). It would therefore seem that in this case, diatom influx figures do provide a reasonable approximation to the palaeo-productivity of Lough Neagh.

What then are the most likely causes of Lough Neagh's enrichment? Archival research has shown that the onset of eutrophication coincided with the Victorian introduction of piped sewerage systems (Patrick, 1988). Instead of being used as land fertilizer and absorbed by the soil system, sewage was discharged as effluent into rivers and thence into the lake itself. Phosphorus inputs have continued to increase during the twentieth century, especially from around AD 1915 when Lough Neagh came to be used for waste disposal by nearby towns, and from the 1950s with the advent of phosphate detergents. The fact that these cultural changes coincided with ecological ones in the lake does not constitute proof that the former caused the latter, but it does represent convincing circumstantial evidence.

Lough Neagh's present-day ecology is the result of cultural eutrophication since the Industrial Revolution. This means that if remedial measures were taken to reduce pollution, there is every prospect of the lake returning to its former less eutrophic state. The ability to recover following disturbance – or elasticity – is one of the features of lakes that make them relatively resilient ecosystems. Once their supply is cut off, pollutants usually become buried in the sediments at the lake bottom or are flushed away down the outflow. This has happened in Lake Washington in the northwest United States, where after public pressure, nutrient-rich waste water was diverted away from the lake after 1967 (Edmondson, 1974). The culturally eutrophied lake is now recovering and recent cores show a declining bulge in the phosphorus content of the lake sediment column. Lake

histories based on sediment cores provide the main way of reconstructing pre-disturbance ‘baseline’ conditions and hence provide a target for restoring polluted aquatic habitats (Battarbee, 1999; Bennion et al., 2011), for example, via the European Water Framework Directive.

Table 7.3 Diatom-inferred changes in total phosphorus (TP) for selected European lakes

	Minimum TP (µg/l)	Date AD	Maximum TP (µg/l)	Date
Augher, Northern Ireland	35	Pre-1900	130	1968
Væng Sø, Denmark	100	1900	160	1987
Langesø, Denmark	135	Pre-1900	225	1975
Marsworth, England	180	1900	486	1983
Talland Sø, Denmark	20–40	Pre-medieval	248	1910

Data from Anderson et al. (1993), Anderson and Odgaard (1994), Bennion (1994) and Bradshaw et al. (2005)

Studies of lacustrine and coastal sediments have shown stories of cultural eutrophication throughout Europe and North America (e.g. Cooper and Brush, 1991; O’Sullivan, 1992; Turner and Rabalais, 1994; Boesch et al., 2001; Bradshaw et al., 2005). Statistical calibration of these data, for example, using weighted averaging techniques, permits quantitative reconstructions to be made of past phosphorus and nitrogen levels (Table 7.3). Anomalously productive ecosystems in regions of otherwise oligotrophic lakes, such as Shagawa Lake, Minnesota, have been revealed as the product of human impact, in this case from mining and sewage waste disposal (Bradbury and Waddington, 1973). The cultural causes of eutrophication are activities associated with industrial and urban society – piped sewage, chemical fertilizers and detergents and toxins such as tributyltin (TBT) and industrial waste. In some cases, these create particularly unpleasant chemical cocktails which can knock out higher-level organisms such as molluscs and fish, with consequences throughout the food chain (Sayer et al., 2006). In non-industrialised regions of the world, pollution by nutrient-rich waste water has been a much less significant problem, although this is changing fast with Third World urbanisation. Eutrophic lakes do exist in the tropics but not always as a result of pollution. Lake George in Uganda, for example, is a highly productive ecosystem with dense crops of blue-green algae and diatoms, but palaeolimnological studies have shown it to have been that way for at least the last 500 years (Viner and Smith, 1973; Haworth, 1977). An attempt to ‘rectify’ this naturally eutrophic lake would be unlikely to succeed!

Acidification and atmospheric pollution

Eutrophication came to be recognized as a major environmental problem during the 1970s, and hard on its heels came the issue of **acidification**. Increased

atmospheric deposition of acids is believed to have had wide-ranging environmental consequences including tree death, accelerated weathering of building facades and damage to human health (Battarbee, 1994). However, probably its most widespread impact, and the one which is discussed here in detail, has been on freshwater ecosystems. In many respects, lake and stream acidification represents the opposite trend to that of eutrophication. Whereas eutrophic waters are murky but nutrient-rich and productive, acidified ones are clear but nutrient-poor and infertile. Acidification starts with the emission of oxides of sulphur (SO_2) and nitrogen (NO_x) into the air from sources such as car exhausts and coal-fired power stations. It has long been known that atmospheric pollution in urban areas leads to 'dry' fallout of sulphur and that this poses an ecological threat, for example, to lichens which are sulphur sensitive. Precisely in order to avoid the problem of local dry fallout of pollutants, high stacks were constructed from which emissions would be carried up and away into the atmosphere. It was hoped that the resulting dilution would be the solution to the pollution!

Once up in the atmosphere, SO_2 and NO_x undergo chemical transformation, involving combination with water vapour to produce dilute sulphuric and nitric acids. This 'acid rain' is carried downwind by atmospheric circulation and is returned to earth as 'wet' deposition in the form of rain or snow. Hydrological processes then lead to the acidified waters entering streams and lakes, with consequences for their plant and animal communities. Some organisms thrive in a strongly acidic environment, but most do not, and there is normally a decline in both total biological activity and in species diversity below pH c.6.0. In the case of 115 southern Swedish lakes, algal species numbers were found to decrease dramatically below a threshold level of pH 5.8 (Almer et al., 1974). The precise ecological mechanisms associated with acidification of freshwaters are only partially understood, but one important aspect is the mobilisation at low pH of aluminium, which is toxic to many higher organisms such as fish. A decline in fish stocks is one of the main characteristics of acidified waters and one that caused Scandinavians and Canadians much anxiety. What made them not just anxious but angry is that the pollution appears to have been exported to them by their neighbours in central or western Europe and the United States, respectively. Acidification is an international problem in which the perpetrator might hope to escape notice by being hundreds of kilometres from the scene – or at least the consequences – of the crime.

This is, of course, a simplified and stylised account. For example, one lake watershed may be experiencing acidification, while another immediately adjacent shows no such sign. Acid precipitation clearly does not have the blanket effect over the landscape that might be expected from atmospheric deposition alone; catchment conditions must also be an integral part of the problem. In particular, the ability of catchment soils and bedrock to neutralise acid precipitation critically influences the pH of the resulting runoff. Buffering capacities are high in areas of basic bedrock such as limestone and low on granite and other acidic rocks. Vulnerability to acidification is therefore greatest where there is a combination of high acid loadings and low catchment buffering capacities. Consequently, any reduction in emissions is most effective if targeted at vulnerable sites, rather than applied 'across the board'. For this, the **critical loads**

concept is widely utilized (Battarbee, 1994).

Amongst the most vulnerable areas where the critical load of acid deposition is most quickly exceeded are the southern portions of the Canadian Shield and Scandinavia which once lay beneath major Pleistocene ice sheets. These areas have thin soils and innumerable lakes, most of which had relatively acidic waters in any case. But how acidic? Some researchers suggested that progressive natural acidification during the course of the Holocene may go a long way to explaining the existence of presently acid lakes. A long-term perspective was therefore required. The history of pH can be measured from lake sediment profiles by analysing their chemical composition or from the hard parts of biological indicators such as invertebrates and chrysophytes. As with eutrophication, one of the most effective palaeo-acidity proxies comes from diatom analysis. Reconstructions of lake acidity have used the pH optima of individual diatom species calibrated statistically by using weighted averaging techniques (Birks et al., 1990). Working at Lake Oresjön in southern Sweden, Ingemar Renberg (1990) used this statistical approach to calculate that there had been a decline of about 2.0 pH units between 12 600 and 2500 years ago. This was a natural change brought about by progressive leaching of catchment soils which had been base-rich at the start of the postglacial, but which later became acidic and peaty. He also found a much more abrupt fall of 1.5 pH units in the top few centimetres of the sediment cores dating to the twentieth century. What is more, this disturbance took Oresjön below pH 5.0 for the first time in its history. It was clear that this and other Scandinavian lakes were not in their ‘natural’ state – their previously stable baseline equilibrium has been seriously disturbed during the last hundred years or so.

Table 7.4 Acidification data for Galloway lakes, Scotland, based on diatom calibration and ²¹⁰Pb dating

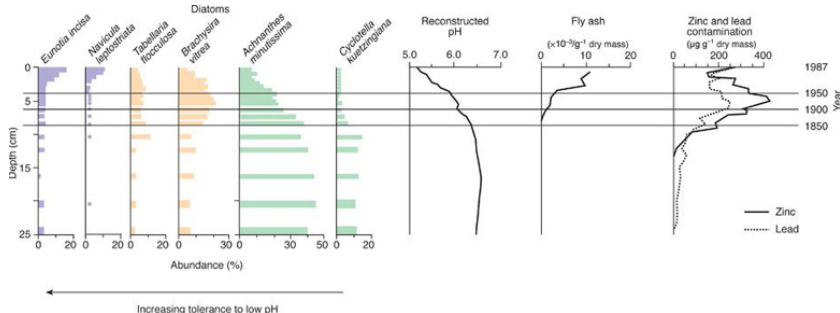
Source: Battarbee (1984). Reproduced with permission of Rick Battarbee.

	Pre-acidification pH	Modern pH (predicted)	Modern pH (observed)	pH decline	Date of tree planting	Date of onset of acidification
Loch Enoch	5.2	4.3	4.4–4.7	0.9	–	1840
Round Loch of Glenhead	5.7	4.7	4.5–5.0	1.0	–	1850
Loch Grannoch	5.6	4.4	4.4–4.9	1.2	1962, 1976	1925
Loch Dee	6.1	5.6	4.9–5.9	0.5	1976	1890

Although the progressive acidification hypothesis could now be ruled out as the sole explanation, this did not make acid precipitation necessarily guilty either. Closer examination was required to establish exactly which human activities have been responsible for recent ecological impact on acid freshwaters. Apart from acid precipitation, another possible anthropogenic cause involved land-use changes in lake watersheds, especially recent afforestation by conifers which can cause increased uptake of nutrients and thereby reduce soil buffering and can also trap and concentrate acid gases. In the Galloway lakes of southwest Scotland, Rick Battarbee, Roger Flower and colleagues from University College London set out to

test these competing hypotheses. They did this by trying to eliminate – in true ‘who dunnit?’ style– the possible culprits (Battarbee et al., 1985). To do this, they considered four lake catchments, two of which had been recently coniferised and two of which were unforested. Diatom analysis of cores from the four lakes showed essentially similar results; all four lakes had been acidified by about 1.0 pH unit from their previous baseline condition, much as at Lake Oresjön (Flower and Battarbee, 1983; Kreiser et al., 1990). Furthermore, ^{210}Pb dating has confirmed that acidification began in the late nineteenth and early twentieth centuries (see Table 7.4), 50 years *before* afforestation began. The planting of conifers may have aggravated the problem, but certainly did not start it. The principal remaining suspect was acid precipitation, an outcome also reached by researchers using comparable methods in eastern North America (Charles et al., 1987; Davis, 1987).

Figure 7.11 pH history of Loch Chon in the Trossachs of central Scotland reconstructed from diatoms, along with the sediment record of atmospheric deposition of fly ash, zinc and lead (adapted from Kreiser et al., 1990).

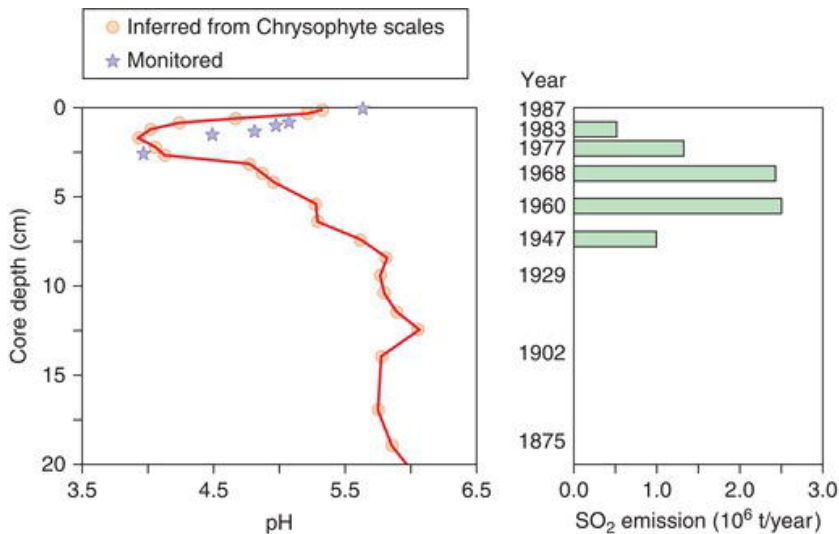


Many other atmospheric pollutants have been recorded stratigraphically in environments ranging from peat profiles through alpine lake sediments to ice cores (e.g. Renberg et al., 1994). Metals such as lead (Pb), zinc (Zn) and copper (Cu) are found at higher levels in acidified lakes, and they have typically increased as pH has fallen (Figure 7.11). While it is unlikely that these trace metals, or the fly ash particles found in lake cores, have themselves significantly contributed to reduced pH, they do give strong support to the link between industrial pollution of the atmosphere and changes in freshwater ecosystems. However, in the Baltic Republic of Estonia, atmospheric pollution has not caused acidification, but the opposite! Northeast Estonia contains one of the world's largest commercially exploited oil shale deposits, and their combustion in major thermal power plants has led to the release of calcium-rich fly ash which has a pH value as high as 12! Much of the surrounding region comprises lakes and peat bogs which are naturally acidic, and their ecology is threatened by this alkaline airborne pollutant. Analysis of peat and lake cores has shown that since the 1950s, acid-loving taxa have been replaced those preferring alkaline waters, amounting to an increase of more than 1 pH unit (Punning, 1994). *Sphagnum* peat has been destroyed and about one-third of the typical bog vegetation has disappeared.

Public pressure, backed by scientific research, has encouraged many

governments and industries in Europe and North America to reduce pollution emissions during the decades since ‘acid rain’ first emerged as an environmental issue, although it remains a major problem in rapidly industrializing countries such as China where air quality can be appallingly bad. Has clean-up brought a tangible improvement in environmental quality in wilderness areas? On the remote ice cap of Greenland, atmospheric deposition of lead (Pb) increased twenty times between c.1800 and 1970 (Murozumi et al., 1969), but lead concentration in Greenland snow profiles has subsequently decreased in response to the decline in use in leaded petrol (Rosman et al., 1993). Similarly, mercury concentrations in lake sediments, which had increased fourfold in the US Midwest since AD 1800, have shown a decline since 1980 (Engstrom and Swain, 1997). Monitoring of acidified lakes has also shown some pH values starting to rise again after decades of decline. Will these acidified lakes be able to recover biologically, and if so, how long will it take? An encouraging illustration of the powers of recovery (or elasticity) of lake ecosystems comes from the Sudbury area of Canada, which has one of the world’s largest smelting plants and the tallest chimney stack in the western hemisphere. Palaeolimnological investigation of 22 out of the more than 7000 lakes in the immediate damage zone around Sudbury had shown a twentieth-century decline in lake pH and increases in aluminium and zinc deposition (Dixit et al., 1995). However, reverse acidification has started to occur in some sites as sulphur emissions have been regulated (Figure 7.12). Chrysophyte scales in a short sediment core from Swan Lake also show a reversal of the previous trend, with a time lag of only few years for biological recovery (Dixit et al., 1989). Higher organisms like brown trout (*Salmo trutta*) and the natterjack toad (*Bufo calamita*) will doubtless take longer to return than this. Nonetheless, it offers persuasive evidence that pollution controls can be followed by environmental improvements within human timescales.

Figure 7.12 Reverse acidification recorded in the sediment profile of Swan Lake, Canada (adapted from Dixit et al., 1989). Right-hand graph shows SO₂ emissions from the nearby Sudbury stack. Reproduced with permission of the National Research Council of Canada.



It is well known that the advent of industrial capitalism has caused the release of pollutants into the environment to shoot upwards. What sediment records have been able to show is, firstly, the widespread nature of their deposition; secondly, how previously stable aquatic ecosystems have often been seriously disturbed in consequence; and thirdly, that many of these systems are self-cleansing once pollutants are reduced below critical load thresholds.

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The environmental future: A Holocene perspective

The history of mankind is a long and diverse series of steps by which he has achieved ecologic dominance.....largely he has prospered by disturbing the natural order.

—Carl Sauer, 1952, pp. 3–4

Some of the steps to humankind's ecological dominance have been described in preceding chapters of this book and are summarised in [Figure 8.1](#). During the late Pleistocene Ice Age, natural agencies – particularly climatic change – played the dominant role in determining the character of environmental systems. The glacial-to-interglacial transition brought dramatic changes to the Earth system, with sea levels rising and biota striving to keep pace with often rapid shifts in climate. The start of the present interglacial was consequently a period of dynamic change and landscape instability, marked in many areas by high rates of erosion and sediment flux (Dearing and Jones, 2003). By the early Holocene, the world's environments and biomes had begun to take on their modern 'natural' form, although this process was not complete until the mid-Holocene in the Arctic and the tropics. The opportunities offered by the Holocene world in turn encouraged new forms of cultural adaptation to emerge, including the domestication of plants and animals.

Relations between nature and culture have been transformed over the course of the Holocene as different human modes of production have emerged or declined. Early human–environment relations, for instance, involved a direct dependence on wild plant and animal resources. Culture was continuous with nature, as it is in contemporary h-f-g groups such as Africa's Bambuti pygmies, for whom the tropical rainforest is a mother and a home, not a wilderness to be tamed. With agriculture, a critical boundary was crossed which led to greater human control over nature and the widespread creation of semi-natural agro-ecosystems. Animals were no longer blood relations but servants. But if peasant farmers and nature formed a separate dualism, they nonetheless constantly interacted to shape each other dialectically. And while hunter-gatherers and peasants are in some respects opposites, at least they share in common an intimate relationship with their natural habitat and a deep respect for nature.

Only under contemporary industrial and post-industrial society has that interdependence been lost and the separation of people and nature been made complete (see [Figure 8.2](#)). The separation between industrial society and the natural environment is well illustrated by the example of acid pollution of freshwaters, whose history has been so effectively revealed by palaeolimnology (Smol, 2008). Even though industrial emissions of sulphur and nitrogen oxides are physically remote from forest and mountain lakes, they have been capable – via long-distance atmospheric pollution – of upsetting the stability of these aquatic ecosystems through acidification. Another powerful illustration, and one critical to our future, is the increasing level of greenhouse gases in the atmosphere brought about by human activities. Concentrations of gases such as carbon dioxide have risen dramatically since the start of the Industrial Revolution, linked particularly to the burning of fossil fuels, and they are responsible for the major share of current climatic warming (Hegerl et al., 2007). Today's period of human-induced changes in the global environment is clearly unique on geological timescales, and this evident most clearly for the Industrial era during the last two centuries. On the other hand, designating the beginning of an epoch dominated by human activities – the Anthropocene – is not so straightforward (see [Technical Box IX](#)). Atmospheric concentrations of CO₂ and CH₄ started rising in mid-Holocene times, plausibly linked to land-use changes associated with early farming societies (see [Technical Box VIII](#)). Wet-rice cultivation and early forest clearance for farmland appear to have led to initially small but cumulative shifts in the global carbon cycle. Unintended human diversion of climate's natural course may therefore go back thousands of years, not just a century or two (Ruddiman, 2005). Similarly, sedimentary records show that in terms of cumulative total, more lead (Pb) polluted Europe's atmosphere during the pre-Industrial era (Roman, medieval, etc.) than has been the case for the last two centuries (Renberg et al., 1994; Martínez-Cortizas et al., 1999). The pivot point in the relationship between human and natural causation of global environmental changes may not therefore always be easy to identify, nor did it always occur as recently as many might think.

Figure 8.1 Summary chart of Holocene environmental and cultural changes.

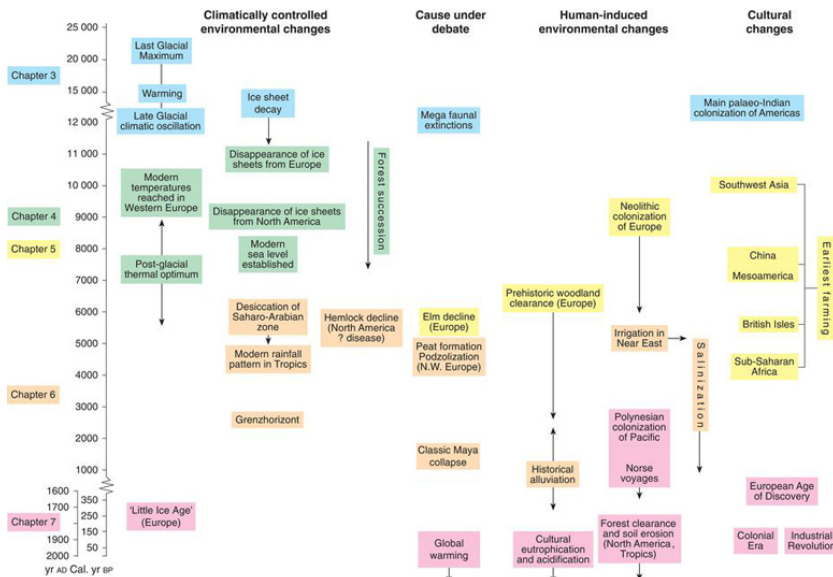
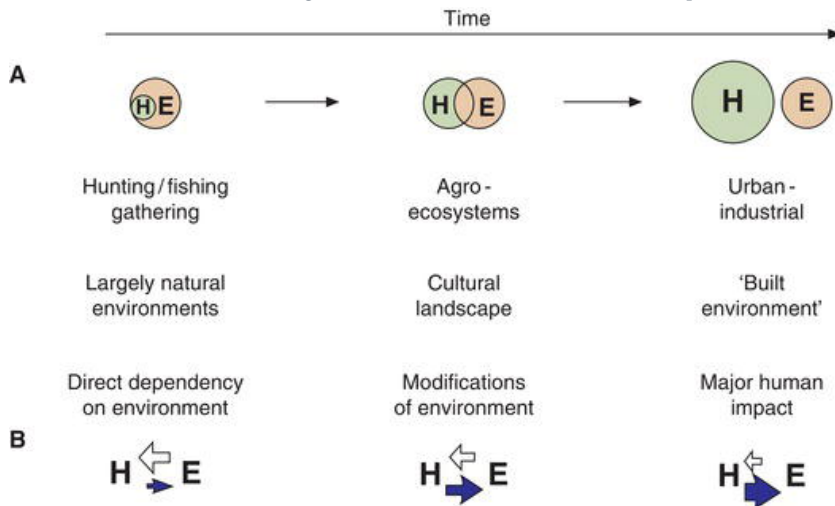


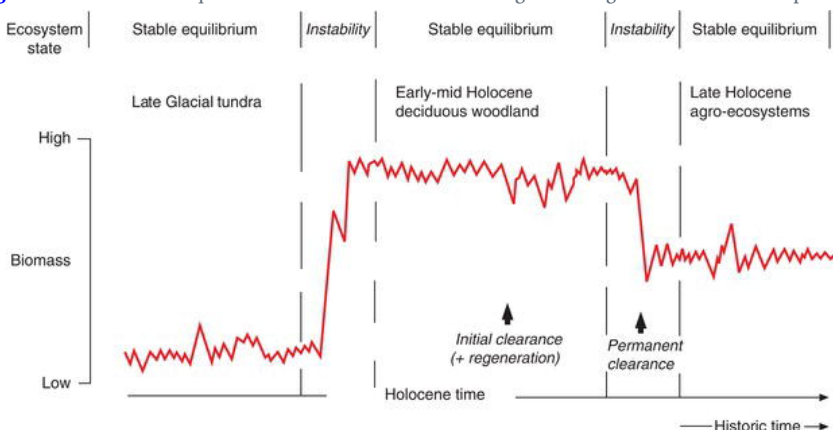
Figure 8.2 The changing relationship between humans (H) and the natural environment (E) over the course of the Holocene, including A nature of interaction and B relative impact.



When viewed over the long term and the large scale, human impact upon the natural world seems to have increased relentlessly, taking us from environmental changes determined by natural forces to those dominated by human actions (Messerli et al., 2000; Dearing et al., 2006; Ellis et al., 2013). In reality, this seemingly inexorable increase in human impact has been formed out of an agglomeration of specific regional histories, each one different. Compare, for example, the history of human–environment relations in southwest Asia and New Zealand. For the former, the impact of Neolithic agriculture on the environment

was being felt even while ice greater than 1 km thick still lay over North America. In contrast, there was no human presence at all in New Zealand and many other Pacific islands for more than 90% of the Holocene. At a regional scale, environmental histories often display marked discontinuities, and at local scales, this is even more true. When Polynesian peoples arrived in New Zealand around thousand years ago, they encountered a truly pristine ecosystem, and first contact was soon followed by forest clearance. In the case of Easter Island (Rapa Nui), this led to complete deforestation (Bahn and Flenley, 1992; Hunt, 2007). In other words, at local and regional scales, it may often be more appropriate to see the changing face of nature as a series of meta-stable equilibrium states than as following a progressively changing trajectory (Oldfield, 1983). According to a meta-stable equilibrium model, prolonged periods of environmental stability were punctuated and separated by much shorter-lived phases of rapid, sometimes catastrophic, change (see [Figure 8.3](#)). These unstable events represent ecological crises or tipping points, during which key thresholds were crossed.

Figure 8.3 Metastable equilibrium model of Holocene ecological change in northwest Europe.



Holocene environmental crises

While palaeoenvironmental records have helped to establish the links between contemporary problems of pollution and land degradation with the emergence of industrial capitalism and colonialism, they also show that environmental crises are by no means the monopoly of modern times. Human-induced ecological problems can be traced back even before the last ice sheets started to wane over Europe and North America, and have involved societies operating under very different modes of production. The catastrophic extinction of most of America's large mammal species around 13 000 years ago would almost certainly not have occurred without the involvement of palaeo-Indian hunters (Barnosky et al., 2004). In Australia, similar extinctions of megafauna followed the first arrival of

aboriginal peoples just before 40 000 years ago and – coupled to increased burning of vegetation – substantially transformed the landscapes of that continent (Miller et al., 2005). Australian pollen records indicate that this human-induced ecosystem shift was as large as any effect of climate change over the last-glacial cycle (Rule et al., 2012). In similar fashion, the formation of blanket peat bogs over much of western Britain and Ireland during the mid-Holocene was inadvertently triggered in many cases by the felling of trees by prehistoric pastoralists and farmers (Charman, 2002; Simmons, 2003).

Nor have environmental crises been uniquely the responsibility of human beings. The catastrophic demise of the North American spruce forest at the start of the Holocene occurred as it was squeezed against the retreating Laurentide ice sheet (Shuman et al., 2002). Other biotic crises such as the decline of hemlock in eastern North America 4800 years ago were caused by natural factors such as pathogen (disease) attack (Bhiry and Fillion, 1996). Since the time of the **LGM**, the Earth has seen some extraordinarily abrupt shifts in climate, particularly at the start of the Holocene, when the planet's climate system switched from a glacial to an interglacial state within less than a century (Alley, 2000). Another tipping point may have been crossed at the end of the African Humid Period around 5500 years ago, as the Sahara–Sahel region dried out (deMenocal et al., 2000). In this case, feedback effects involving changes to the regional water cycle and land-surface albedo may have caused a non-linear climate response, which led to a new, drier meta-stable equilibrium climate state.

Global temperatures have been remarkably stable for the last 10 000 years, compared to the huge swings between cold and warmth which characterized the late Pleistocene. Nonetheless, climatic oscillations in the form of high-latitude cooling events and low- to mid-latitude droughts have occurred every 1500–3000 years during the Holocene (Gasse and van Campo, 1994; Bond et al., 1997; Mayewski et al., 2004; Wanner et al., 2008). Not surprisingly, climatic change had a marked effect on the stability of some environmental systems and human societies (e.g. Mainwaring et al., 2010). What is equally striking is that societal responses to past climatic crises were far from uniform, even in the same region. At the end of the early Bronze Age around 4200 years ago, for instance, mega-drought affected the eastern Mediterranean region (Weiss et al., 1993; Cullen et al., 2000). While this prompted some societies to build ever bigger temples to appease the gods, other groups instead responded by developing new water systems for irrigation and domestic supply (Rosen, 2007). Needless to say, the latter proved to be more resilient than the former.

What circumstances produced these environmental crises, and what lessons can be learnt from them to help avert them in future? The main relevant factors, listed in **Table 8.1**, may be condensed down to ecosystem resilience on the one hand and the type of impact on the other (Redman and Kinzig, 2003). Some ecosystems are robust, like grassland environments, or are 'elastic' in their recovery, like lakes. However, others, such as the limestone plateau of the Burren in western Ireland, had a low threshold for change coupled with an inability to recover following disturbance. Whether the changes in a particular environmental system are reversible or not frequently depends on the maintenance of a soil suitable for plant growth. In the case of the Burren, the soil erosion that followed prehistoric and early historic forest clearance was so severe that much of the

landscape was left as bare limestone. The preservation of the soil is therefore a vital factor in the maintenance of ecological habitats as a whole. This is one reason why soil degradation has been and still is such a critical environmental issue, whether in the Burren 3000 years ago, Oklahoma in the 1930s or in Lesotho today.

Table 8.1 Factors determining ecosystem response to disturbance

Ecosystem resilience	Robust ----- (e.g. grassland)	Fragile (e.g. island ecosystems)
Ecosystem recovery	Elastic ----- (e.g. lakes)	Irreversible (e.g. badland erosion)
Origin of disturbance	Internal/gradual ----- (e.g. soil leaching)	External/abrupt (e.g. first human arrivals)
Duration of disturbance	Short lived ----- (e.g. volcanic eruption)	Sustained (e.g. marine transgression)
Continuity of disturbance	Episodic ----- (e.g. wildfire events)	Progressive (e.g. eutrophication)
Overall ecosystem response	Maintenance of stability/ recovery	Instability/degradation/collapse

The Burren needed only a small ‘push’ for it to cross the threshold into environmental instability, and the same is true of the history of other inherently fragile ecosystems like those parts of upland Britain today blanketed by peat bog. With ecological shifts such as peat initiation, it is probably wrong to talk of human and natural causation separately, for human impact would have been ineffective without suitable environmental preconditions (Head, 2008). In fact, most Holocene ecological crises have involved several factors operating in combination. The European elm decline, for instance, is most plausibly explained by the combined impact of Neolithic farming and a pathogen outbreak (Parker et al., 2002). Whereas the impact of either factor on its own might have been absorbed by the elm tree population, the combination of both factors produced environmental stresses which could not be withstood. Significantly, in North America, the hemlock tree (*Tsuga*) did eventually recover to a status similar to that before its decline – while in Europe the elm did not. If there is a moral here for contemporary environmental problems, it is that human disturbance may not on its own produce ecosystem breakdown but that it will weaken the resistance of ecosystems to other environmental impacts.

Another very important influence on Holocene environmental crises lies in the origin and nature of disturbance. Abrupt, external impacts have usually had much more dramatic consequences than gradual, internally generated changes. This applies to disturbances of natural origin – for example, abrupt shifts in climate – just as it does to cultural ones, such as the arrival of the Polynesians in New Zealand or Easter Island. The degree of abruptness goes a long way towards explaining why late Pleistocene mammalian extinctions were so severe in the Americas but were hardly felt at all in Africa. Hominins had co-existed with the megafauna of the African savannas for millions of years, and the two had adapted

to one other and co-evolved. By contrast, *Homo sapiens* only arrived in the New World at the end of the Pleistocene, expanded rapidly and was unfamiliar to the unsuspecting indigenous animal population. In other words, human disturbance increased gradually in Africa, giving the biota time to adjust, whereas in the Americas, the impact came abruptly and catastrophically. Impact is also likely to have been greater if sustained, rather than being a short-lived perturbation. The largest single abrupt climatic event of the Holocene occurred around 8200 years ago was of the latter type, linked to the final demise of the North American ice sheet. This prompted cool and/or dry climatic conditions that lasted for around a century, but then returned back to pre-disturbance conditions (Alley and Ágústssdóttir, 2005; Daley et al., 2011). As a result, the long-term environmental consequences of the 8.2 ka BP climate event were rarely significant; in the centuries that followed, for example, the Sahara re-greened and vegetation re-adjusted (Lamb et al., 1995; Hoelzmann et al., 2004). The key tipping point for this regional dryland system was crossed only some three millennia later, when underlying boundary conditions had altered linked to changed orbital forcing.

Environmental conservation and Holocene environmental history

Palaeo-science helps in understanding the trajectory of an environmental system and hence aid in its management (Anderson et al., 2006) and in quantifying past changes in ecosystem services, such as loss of clean drinking water due to pollution of natural sources (Dearing et al., 2012). It can help to establish, for example, whether present levels of rainfall and water-resource use are sustainable over the long term. At Naivasha in Kenya, for example, colonial settlement at the turn of the twentieth century happened to coincide with a period of unusually wet climate and high water levels (Verschuren et al., 1999), which gave a false impression of the area's agrarian potential. Past records can further help in the identification and designation of ecosystems which have survived essentially undisturbed through the Holocene. For example, Lake Baikal in Siberia is the world's largest and deepest freshwater lake and has a rich and unique biota, even including a freshwater seal. Analysis of short cores from the lake has shown that while the uppermost sediments have begun to be contaminated with pollutants like lead, so far there have been no detectable consequences of human disturbance on the ecosystem as a whole (Flower et al., 1995). In other ecosystems which have already been impacted, proxy records can allow their pre-disturbance baseline condition to be identified, which can then help in setting restoration management targets (Gillson and Willis, 2004; Willis et al., 2007).

On the other hand, palaeo-studies show that environmental systems did not always have a single stable baseline state. An instructive example comes from

Sweden where Ingemar Renberg carried out detailed reconstructions of lake histories, particularly diatom-inferred pH. In Lilla Öresjön, there was a marked increase in lake acidity caused by atmospheric pollution, with pH declining to a minimum of 4.6 in 1986 (Renberg, 1990). Prior to AD 1900, diatom-based reconstructions indicate stable pH conditions around 6.0, which would appear to provide a good target for lake water quality restoration. However, further down in the lake sediment cores, diatoms and inferred lake water acidity changed again. Prior to 2300 Cal. yr BP (300 BC), the pH of Lilla Öresjön was between 5.0 and 6.0, in other words closer to the values brought about by industrial-age pollution and acid deposition! The cause of this earlier change was the arrival of the first farmers, who replaced h-f-g populations and converted forest to mixed land use, and led to anthropogenic alkalisation of the lake ecosystem (Renberg et al., 1993). Complex environmental histories such as this create a challenge for environmental managers. As a restoration target, should they use the most natural Holocene environmental state, in this case prior to Iron Age times, or the one which existed for most of the last two millennia? In fact, for management programmes such as the European Water Framework Directive, it may not be sensible to use any past ecosystem state too rigidly as a restoration target, because climate change is likely to change conditions in the future (Bennion et al., 2011). Environmental systems have always been dynamic, and they will continue to respond dynamically to future conditions, so that their management should adjust target states accordingly.

In relatively undisturbed environments, such as ancient woodland, genetic and other characteristics are retained which are absent in more disturbed areas of secondary woodland. Preservation of the gene pool is especially important where little of the original vegetation remains intact. For instance, only 0.2% of Ireland lies under 'native' woodland – equivalent to the area occupied by the city of Dublin – and most of this is secondary. Pollen studies have shown several western Irish 'ancient woods' in fact to be of recent origin and that they are still evolving in terms of their species composition (Hannon and Bradshaw, 1989; Mitchell, 1990). The recognition that wildfire is a long-standing part of many ecosystems is another way in which ecological history has helped to influence environmental management strategies. Fires can act to release and make available mineral nutrients, regulate dry matter accumulation and control forest insects and parasites. Misguided fire suppression by forest managers has sometimes had serious adverse effects on forest ecology and increased the risk of a major, uncontrollable conflagration, such as occurred in Yellowstone National Park in 1988. Fire scars on dated tree rings and charcoal peaks in lake sediments can help to establish the 'natural' magnitude and frequency of forest fires during the period before written observations and before twentieth-century fire suppression policies were applied (Swetnam, 1993; Millsaugh and Whitlock, 1995; Larsen, 1996; Conedera et al., 2009).

The example of fire ecology illustrates how far removed from nature urban, industrial society became in its philosophy of environmental conservation. According to this philosophy, the natural environment should be preserved rather than being allowed to change dynamically. 'Leave it as it is', said US President Teddy Roosevelt at the Colorado Grand Canyon in 1903. However, very little in the natural world is stable when viewed over the timescale of centuries and

millennia, and nature should not be made sacred in the name of conservation. Individual species have risen and fallen in importance during the Holocene, making the most of opportunities presented to them (Willis and Birks, 2005). For example, beech was relatively unimportant in the British Isles until agricultural clearance provided gaps in the forest for it to expand late in the Holocene, and it was absent in Ireland until the last millennium (Giesecke et al., 2007). By contrast, pine was common in Ireland for most of the Holocene but died out after 2000 years ago (Bradshaw and Browne, 1987). Trying to use species conservation to stop the ecological clock at AD 2000 would be as artificial as designating one of these trees to be a native species but the other not.

The national park idea, developed in North America with Teddy Roosevelt's support, and subsequently exported to other parts of the world, reflects the philosophy that the natural environment should be preserved as sacrosanct rather than utilised. The International Union for the Conservation of Nature and Natural Resources (IUCN) describes a national park as 'one or several ecosystems not materially altered by human exploitation or occupation...and where the highest authority has taken steps to prevent or eliminate as soon as possible exploitation or occupation in the area'. In reality, many ecosystems are far from being wholly 'natural' and instead owe their distinctive character to particular patterns of land use or other human actions. For example, Garaet el Ichkeul in Tunisia is a shallow coastal lake which is one of the most important wetland sites for migratory wildfowl around the Mediterranean, but its present ecological state only came into being in the nineteenth century when French engineers built a ship canal into an adjoining lake and inadvertently allowed the inflow of saltwater (Stevenson and Battarbee, 1991). Without this unintended human impact, Ichkeul would not have its present high conservation value! Similarly, many of our most valued 'natural' landscapes – such as the 'wilderness' of Dartmoor and the scrub oak–pine barrens of New England – are plagioclimax ecosystems created and partly maintained by traditional land-use practices such as grazing and burning. The idea that the wilderness should be free of people living and working there has less to do with successful ecosystem functioning than it has with the romantic ideal of a *primaevae* nature independent of people. Palaeoecology shows that ideal to be a myth. In all but a few cases, humans have been an integral part of ecosystems since present-day climates and environments became established early in the Holocene. For most ecosystems, it is therefore effectively impossible to study environmental history separate from cultural history and vice versa.

The 'best leave alone' conservation philosophy makes even less sense in those parts of the world where peasant farming, pastoralism or h-f-g today remain strongly embedded. Societies operating under these modes of production have, over time, slowly but progressively established a detailed knowledge of their local habitats. This indigenous environmental knowledge, for example, of tree species and their different uses, is geographically specific and cannot be built up overnight. In these cases, conservation needs to include cultural traditions as much as it does flora and fauna. The real issue is not one of survival and preservation but of incorporating traditional cultural ecologies into future development. Building on such traditional forms of culture–environment relations from the bottom up is more likely to be successful imposed, alien technological 'solutions'. A good example is provided by wetland horticultural systems, which

have a long antiquity in the tropics (e.g. Golson, 1977; Darch, 1988). What is more, palaeoenvironmental research has shown these to be ecologically sustainable, nutrient-conserving systems, which do not normally cause long-term soil degradation. Small-scale garden horticulture remains important in many parts of the tropics and offers an alternative to conventional large-scale irrigation schemes.

One of the most pressing of the world's contemporary conservation issues is the continuing destruction of tropical forests, especially in Amazonia. Sadly the present pattern of exploitation shows few signs that the lessons of environmental history have been learnt and acted upon, if we compare recent developments with the indicators listed in [Table 8.1](#). Prior to European conquest and diseases, evidence from geoglyph earthwork structures and *Terra Preta* soils indicate that the Amazon probably supported several million people, apparently without compromising rainforest biodiversity (Heckenberger et al., 2003, 2007; Mann, 2005; Arroyo-Kalin, 2012). By contrast, clear-felling is pushing the rainforest beyond the threshold of ecological stability and leading to severe soil erosion, hence making forest regeneration difficult or even impossible. To judge by examples from the environmental past, abrupt external impact on a fragile environment has not been conducive to ecological stability. Even so, human-induced rainforest clearance has so far been no worse than the naturally caused Ice Age deforestation of the Earth, and during the Holocene, the forests – tropical and temperate – were able to reclaim most of that area within a thousand years (Mayle et al., 2004; Metcalfe and Nash, 2012). But can we afford to wait that long?

Lest this book end pessimistically, let it be noted that environmental history offers us not only lessons but also hope for the future. We are a remarkably ingenious species, and past examples tell us that environmental adversity has prompted innovation as often as societal collapse (Butzer, 2012). At some times and in some places, we have adapted nature beautifully and harmoniously, working together rather than seeking crude exploitation. However, we need to learn that most of nature's timescales of change are not the same as those which we experience individually as human beings (Roberts, 2011; Caseldine, 2012). The wisdom of our decisions about environmental resource use and carbon emissions will be judged over the coming decades and centuries, not at the end of a four-year presidential mandate. Harmonizing the timescales of human decision-making with those of global environmental change is one of the greatest challenges facing us. Even if it appears that we can get away with behaving selfishly in the short term, in the long run our fate – one way or the other – will be inextricably bound to that of the planet as whole.

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Appendix: Calibration table for radiocarbon ages

¹⁴ C age (yr BP)	Calendar age (Cal. yr BP) intercepts and range	Calendar age (Cal. yr BP)	¹⁴ C age equivalent (uncalibrated yr BP)
500	519	500	430
1000	928	1000	1120
1500	1354	1500	1600
2000	1938	2000	2060
2500	(2709) – 2581 – (2509)	2500	2450
3000	(3205) – 3187 – (3169)	3000	2900
3500	(3817) – 3792 – (3725)	3500	3320
4000	4436	4000	3670
4500	(5246) – 5122 – (5056)	4500	4010
5000	5731	5000	4430
5500	6294	5500	4790
6000	(6854) – 6821 – (6814)	6000	5260
6500	7384	6500	5720
7000	7787	7000	6130
7500	(8313) – 8220 – (8218)	7500	6700
8000	(8950) – 8784 – (8767)	8000	7260
8500	(9480) – 9474 – (9456)	8500	7760
9000	9979	9000	8110
9500	(10 537) – 10 513 – (10 482)	9500	8560
10 000	(11 800) – 11 700 ^a – (11 200)	10 000	9040
11 000	12 917	11 000	9800
12 000	13 992	12 000	10 230
13 000	15 437	13 000	11 090
14 000	16 792	14 000	12 000
15 000	17 916	15 000	12 740
16 000	18 876	16 000	13 380
17 000	20 105	17 000	14 180
18 000	21 484	18 000	15 070
		19 000	16 120
		20 000	16 920
		21 000	17 630
		22 000	18 410

Source: OxCal.14 L (except for^a) (<http://c14.arch.ox.ac.uk/embed.php?File=oxcal.html>)

Glossary

These words appear in **bold blue typeface** in the text of the book.

Acidification The process by which soils or freshwaters become more acid, i.e. come to contain a higher concentration of hydrogen ions. This may be natural or cultural in origin

Allochthonous Materials deriving from, or processes occurring, outside the system, such as a lake; opposite of **autochthonous**

AMS or Accelerator Mass Spectrometer Dating: a form of **radiocarbon** dating which requires very small sample sizes

Anaerobic (or anoxic) Without oxygen

Anthropocene The most recent time period characterized by global-scale human impacts on the Earth system, normally coinciding with the Industrial era

Autochthonous Materials produced *in situ*, or processes occurring within a system, opposite of **allochthonous**

Biogenic Materials of organic origin

Clovis culture The first well-attested hunter-gatherers known in North America around 13 ka BP. Archaeologically, their material culture has characteristic stone projectile points

Coleoptera Beetles. Their hard exoskeletons made of chitin can become disaggregated after death into different parts (head, thorax, etc.)

Correlation The matching of strata or events of similar age between one place and another; e.g. correlating sediment cores within a lake

Critical load The highest influx of polluting compounds that will not cause long-term harmful effects on ecosystems

Dendrochronology Dating by counting the annual growth rings in trees and old timbers

Diatoms A type of unicellular algae, whose resistant cell wall is preserved after death. Diatom analysis of lake muds forms an important part of palaeolimnology

Domestication The process whereby wild plants are taken into cultivation and wild animals tamed. This usually involves sowing as well as harvesting of plants, tilling of soil by hoe or plough and controlled breeding of livestock

Dystrophic Lakes and streams which are themselves infertile, but contain

dissolved humic substances of allochthonous origin, giving a characteristic brown colour to the water

ENSO An abbreviation for the quasi-periodic climatic oscillation known as El Niño–Southern Oscillation. This is an index of the atmospheric pressure gradient across the South Pacific. Every 5 to 10 years, it switches from its normal mode, causing heavy rainfall in areas – like the Peruvian desert – which are normally dry, and droughts on the western side of the Pacific. Flood–drought cycles believed to be related to ENSO have been identified in many parts of the tropics and subtropics

Environmental Archaeology The study of plant and animal remains and sediments from archaeological sites in order to reconstruct their economy and environment

Equifinality A term used when a similar end result could have been produced by more than one different cause; for example, the mid-Holocene elm decline could have resulted from disease, from human impact or from climate change

Ergodic The substitution of space for time, or vice versa. The woodlands of eastern North America encountered by the first European settlers, for example, have sometimes been thought of as resembling the original, preagricultural landscapes of northwest Europe and might represent a suitable analogue for them

Eustatic The global component of sea-level change, primarily involving the interchange of water between land ice and the sea. As ice sheets melt, the ocean volume increases and sea levels rise

Eutrophic Ecosystems such as lakes and soils which are rich in nutrients and have a high primary productivity. In hyper-eutrophic environments, species diversity decreases and predatory organisms decline

Eutrophication The process of nutrient enrichment, which may be natural or cultural in origin

Feral Domestic plants or animals which have escaped and returned to the wild

Flandrian Name sometimes given to the Holocene in the British Isles

Geo-archaeology The study of sediments at, and adjacent to, archaeological sites in order to reconstruct local environmental histories

Grenzhorizont The boundary horizon in northwest European raised mires marking a change from dark, humified peat to light, less well-humified peat. It has been interpreted as marking a shift from a drier (sub-Boreal) to a wetter (sub-Atlantic) climate at c.2600 yr BP, although radiocarbon dating has shown the horizon to be time-transgressive rather than synchronous across Europe

Holocene The postglacial epoch, is conventionally – if arbitrarily – dated as beginning 10 000 ¹⁴C (or about 11 700 calendar) years ago. Holocene, meaning ‘wholly recent’, was introduced as a term by Gervais in 1869, and accepted by the International Geological Congress in Portugal in 1885

Hominin The group consisting of anatomically modern humans (AMH), extinct human species and all our immediate ancestors, including – but not restricted to – members of the genus *Homo*

Hydrosere A succession from open water to mire and bog at the edge of a lake

Interstadial A warm climatic period of shorter duration than true interglacial, e.g. Windermere interstadial (15–13 ka BP)

Isostatic That component of sea-level change involving compression loading and unloading of the Earth's crust, most importantly by ice. Glacio-isostatic pressure beneath ice sheets depressed parts of Canada and Scandinavia by over 1 km at glacial maxima. Since the LGM, these and adjacent areas have been subject to Glacio-isostatic adjustment (GIA)

Isotopes Different forms of the same chemical element, defined by the number of neutrons present. Unstable isotopes (e.g. ^{14}C) experience radioactive 'decay' and can be used for dating purposes. Stable isotopes provide information on past temperatures and other former environmental conditions

ka Thousand years

Landnám A localised phase of anthropogenic woodland clearance, usually for agriculture and often followed by abandonment and regeneration of secondary woodland

Last Glacial Maximum (LGM) The last time when global ice cover was at its greatest extent, around 20 000 years ago

Late-Glacial (or terminal Pleistocene): the final part of the last glacial stage, dated to between about 15 and 11.7 ka BP

Little Ice Age (LIA) A period of generally colder climate which started around AD 1300 and ended c.1850; it is best recorded in northern and central Europe, but was probably global in extent

Loess Windblown silt, commonly originating from deflation of sediments around former ice sheets. In northern China, loess can be more than 300 m thick

Magnetic susceptibility An easily measured parameter which gives an indication of the 'magnetisability' of sediment, determined by the abundance of ferromagnetic minerals. It is particularly useful in studies of lake sediment core correlation.

Mesocratic The second early-temperate substage of northern European interglacials, when mixed deciduous woodland and brown earth soils became established

Mesolithic The Middle Stone Age period of the terminal Pleistocene and/or early Holocene, whose cultures were reliant on broad-spectrum h-f-g economies

Minerogenic Clastic sediments which comprise individual mineral particles and are usually classified according to grain size (e.g. sand, clay)

MIS Marine isotope stage. Based on the oxygen isotope record from deep-sea sediment cores, this is the main way of dividing up the Quaternary timescales. Warm climatic stages have odd numbers (1, 3, 5, etc.) and cold stages even-numbered ones (2, 4, 6, etc.), counting back from the present day, with the Holocene being MIS 1

Mollusca Small invertebrate animals with an external shell made of calcium carbonate that is well preserved in alkaline soils and sediments

Multi-proxy The use of multiple indicators (or proxies) of past environmental change, often from a single site

Neolithic The New Stone Age period characterised by the adoption of agriculture, village life and use of pottery

Oligocratic The third, or declining, late-temperate substage of northern European interglacials, when soil leaching led to podzolisation

Oligotrophic Lakes, soils or other ecosystems which are nutrient poor and which consequently experience low levels of primary productivity by photosynthesizing green plants or algae

Ombrotrophic Ecosystems, e.g. raised mires, which are dependent solely on atmospheric sources for their supply of water and nutrients

Ontogeny Individual development through time to maturity, e.g. of lake ecosystem

Palaeoecology The reconstruction and study of past ecosystems, including the relations between organisms and their environments

Palaeolimnology The study of lake history, most importantly from evidence preserved in its bottom sediments

Palaeolithic The Old Stone Age period whose Pleistocene cultures ranged from the earliest tool makers at Olduvai in East Africa to the Ice Age hunters of Lascaux

Palynology The study of pollen grains and spores, both at the present day (e.g. for allergy sufferers) and for the past in order to reconstruct former vegetation

Phytoliths Microscopic silica structures formed by plants, especially common in dry and tropical regions

Plagioclimax The end point of a deflected ecosystem succession which is maintained anthropogenically

Pleistocene That period of time, including both glacial and interglacial stages, between 2.6 million and 11 700 years ago. The Late Pleistocene covers the last interglacial-glacial cycle up to the beginning of the Holocene

Pluvial A former period of greater wetness, e.g. one of high lake levels, usually, but not always, due to an absolute increase in rainfall

Protocratic The initial, pre-temperate substage of northern European

interglacials, typically characterised by raw, unleached soils and associated with rapid immigration of pioneer trees such as birch

Quaternary The current period of geological time, which comprises the Pleistocene and Holocene and which began about 2.6 million years ago. It is broadly corresponds to the Last Ice Age

Radiocarbon dating A radiometric method of establishing the age of organic materials up to c.40 000 years old. It is amongst the most widely applied of all Holocene dating techniques

Radiometric dating Those dating techniques based on the principle that radioactive decay of unstable isotopes is time dependent

Sclerophyll(ous) Woody vegetation with hard, evergreen leaves designed to reduce water loss. Often prone to fire

Sedentism Living in one main place year-round (as opposed to nomadism)

Spectral analysis A statistical technique that enables regular cycles but of different length to be identified in a compound time series; e.g. astronomical cycles

Speleothems Cave carbonates such as stalagmites and flowstones

Stadial A cold climatic period of shorter duration than a true glacial phase, e.g. Younger Dryas event (13–11.7 ka BP)

Synchronous Events that happen simultaneously

Telocratic The final, post-temperate substage of northern European interglacials, with declining temperatures, and heathland or open conifer woodland

Tephrochronology The study of volcanic ash layers (tephra). The deposition of tephra following the eruption of a volcano is effectively a synchronous event, and the resulting ash layer provides a marker horizon which can be correlated from one site to another

Time-transgressive Processes, such as the spread of a disease, which started earlier in some places and ended later elsewhere

Uniformitarianism A basic principle of historical geology and palaeoecology, which, in the sense of ‘actualism’ implies that past events can be explained in terms of present-day processes and relationships. Uniformitarianism has also – confusingly – been used in a second sense to mean no significant change in the *rate* at which processes take place (or gradualism)

Varves Annually layered sediments in lakes or the oceans

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Younger Dryas stadial (12,900–11,700 Cal. yr BP)

Dystrophic

Easter Island (Rapa Nui)

Ecology

competition

cultural/human

ecological crises

historical

succession, (*see also* Hydrosere)

Ecosystems

agro-

elasticity of

island

lake-catchment (watershed)

mediterranean-type

response to disturbance

services

Eemian *see* interglacial

Egypt

Elk-moose (*Alces*)

Elm (*Ulmus*)

decline

El Niño *see* ENSO

England, *see also* Chalk downland; Lakes; Mesolithic

Bronze Age landscape and origins of moorland on Dartmoor

pollen diagram of land-use history from Rismoor

pollen record of vegetation from Hockham Mere, East Anglia

ENSO

Epilimnion

Equifinality

Ergodic

Erosion *see* Soils

Euphrates (river)

European

Age of Discovery

Pollen Database (EPD)

Water Framework Directive

- Eustatic
- Eutrophic
 - eutrophication
- Evolution, *see also* Ontogeny
- Extinctions, *see also* Moa bird
 - megafaunal
- Extra-terrestrial impacts

- Fairbanks, R.G.
- Farming *see* Agriculture
- Feral
- Feudal
- Fir (*Abies*)
- Fire
- Firn (or snow) line
- Fish
 - fishing economies
 - kills of
- Flandrian
- Flenley, J.R.
- Fluvial (or riverine), *see also* Alluvial; Palaeohydrology
- Food
 - procurement and diet
 - production
- Foraminifera
- Forest, *see also* Boreal; Deforestation; Tropical rainforest; Woodland
 - fallow
- Fowler, P.J.
- France
- Franchthi Cave, Greece
- Fuller, D.Q.

- Garriga
- Gazelle (*Gazella dorcas*)
- Geo-archaeology
- Geomorphology
- Glacial *see also* Alps; moraines; till
 - climate
 - period
- Glaciers
 - and climate

Glacken, C.J.
Goat (*Capra*)
Godwin, H.
Golson, J.
Goosefoot *see* Chenopods
Goudie, A.S
Grasses
Grassland
Greece, *see also* Franchthi Cave, Greece
Greenland, *see also* Viking
 ice cores
Grenzhorizont

Harlan, J.R.
Harris, D.R.
Hazel (*Corylus*)
Heathland
 mallee
Heinrich events
Helbaek, H.
Hemlock (*Tsuga*) decline
Herbs (or forbs)
H-f-g (hunting-fishing-gathering)
Higgs, E.S.
Hippopotamus
Hockey-stick curve
Holly (*Ilex aquifolium*)
Holocene (or post-glacial) epoch
Hominin
Hornbeam (*Carpinus*)
Horse (*Equus*)
 domestic (*E. caballus*)
 wild (*E. ferus*)
Horticulture
Humus
Hunting-fishing-gathering *see* H-f-g
Huntley, B.
Hutchinson, G.E.
Hydraulic civilization
Hydrosere
Hypolimnion

Iberia

Ice

cores

sheets

Fennoscandian

Laurentide

Iceland

Incremental dating methods, *see also* Dendrochronology

lichen and lichenometry

Industrial revolution, *see also* Pollution

consequences of

Indus valley civilisation

Interglacial

cycle in northwest Europe

last (Eemian/Ipswichian)

Interstadial

IP₂₅

Ipswichian *see* Interglacial

Ireland

evolution of Lough Neagh

history of the Burren

Neolithic clearance at Ballyscullion

Iron (Fe)

Iron Age, *see also* Celtic

Irrigation

Island,

ecosystems

Isostatic

Isotopes, *see also* Radiometric dating

Iversen, J.

Ivy (*Hedera helix*)

Japan

Juniper (*Juniperus*)

Kutzbach, John

Lakes *see also* Sweden

Abhé (Ethiopia)

Baikal (Siberia)

Bosumtwi (Ghana)

- Chad
- Crawford Lake (Canada)
- Erie and Frains Lake (*see* United States)
- Holzmaar (Germany)
- Lake District (England)
- level changes
- Lough Neagh (*see* Ireland)
- Malawi
- Mungo (*see* Australia)
- Naivasha (Kenya)
- Nar (Turkey)
- Oresjön and Havgårdssjön (*see* Sweden)
- sediments
- Suigetsu (Japan)
- Washington
- water balance, (*see also* Pluvial)
- watershed (*see* Ecosystems)
- Windermere
- Zeribar
- Ziway-Shala (Ethiopia)
- Laminations *see* Varves
- Landnám
- Land snails *see* Mollusca
- Land-use history
- Late-Glacial (or terminal Pleistocene) stage (16,000–11,700 Cal. yr BP)
- Laurentide
- Lead (Pb)
 - lead-210 dating
- Libby, W.
- Lichen and lichenometry
- Lime/linden tree (*Tilia*)
- Limnology
- Ling heather (*Calluna*)
- Little Ice Age (LIA)
- Livingstone, D.
- Loch Lomond (or Younger Dryas) stadial
- Loess
- Luminescence (or optical) dating
 - optically-stimulated luminescence (OSL)
 - thermoluminescence (TL)
- Macchia
- Mackereth, John
- Macrofossils

- animal
- plant
- Madeira
- Magnetic susceptibility
- Maize (*Zea mays*)
- Mammoth (*Mammuthus*)
- Manioc (*Manihot*)
- Mann, Mike
- Maori settlement of New Zealand
- Mazama, Mount (volcano)
- McAndrews, 'Jock'
- McIntyre, Andrew
- Mediaeval
- Mediterranean basin, *see also* Ecosystems
 - history of vegetation
- Megafauna *see* extinctions
- MesoAmerica, early agriculture in, *see also* Mexico
- Mesocratic (or early-temperate) substage
- Mesolithic
 - settlement in North England
- Mesopotamia
- Metastable equilibrium
- Methane (CH₄)
- Mexico
- Middle Stone Age *see* Mesolithic
- Migration
 - rates
- Milankovitch, Milutin, *see also* Orbital cycles
- Millet
 - bulrush or pearl (*Pennisetum*)
 - chinese (*Panicum*)
 - finger (*Eleusine*)
 - foxtail (*Setaria*)
- Minerogenic
- Mires *see* Peat
- Mistletoe (*Viscum album*)
- Moa bird (*Dinornis*)
- Mollusca
 - edible
 - land snails
 - marine
 - radiocarbon dating of

Moore, Peter
Moraines
Mugwort *see* Artemisia
Multi-proxy approach

N *see* Nitrogen (N)
Na *see* Sodium (Na)
Natufian culture
 site of Ain Mallaha, Jordon valley
Neanderthal
Neolithic
 agricultural revolution
 Jomon (Japan)
 Saharan (*see also* Aqualithic culture)
 spread across Europe
Netherlands
 geomorphic history of River Mark
Nettle (*Urtica*)
New Stone Age *see* Neolithic
New Zealand, *see also* Maori settlement of New Zealand
Nicholson, Sharon
Nile *see* Palaeohydrology
Nitrogen (N)
Nomadism *see also* Pastoralism
Norman period in Britain
Nutrients
 enrichment (*see* Eutrophic, eutrophication)

Oak (*Quercus*)
 evergreen
Oasis hypothesis of early agriculture
Oldfield, Frank
Oligocratic (or late-temperate) substage
Oligotrophic
Olive (*Olea europaea*)
Ombrotrophic (or ombrogenous), *see also* Peat
Ontogeny
Opportunism
Orbital cycles
Ostracods
Ötzi - the ice man

Oxygen, *see also* Anaerobic
isotope analysis

Pacific Ocean *see also* Easter Island; Polynesian colonization of, Pacific
Packrat (*Neotoma*) middens

Palaeoceanography

Palaeoclimatology

Palaeoecology

Palaeohydrology

of Lake Missoula and Channelled Scabland

of Mississippi River

of River Nile

Palaeohydrology of river Nile

Palaeolimnology

Palaeolithic

Palaeomagnetic dating

Palaeosols *see* Soils

Palynology *see* Pollen, analysis

Papua-New Guinea

Kuk swamp and early agriculture

soil erosion history

Pastoralism

Pathogen *see* disease

Peasant farming

Peat

bogs and mires

formation (paludification)

stratigraphy

Pennington, Winifred

Periglacial

Phosphorus

Pig (*Sus*)

Pine (*Pinus*)

bristlecone (*P. longaeva*, *P. aristata*)

Irish extinction of

jack (*P. banksiana*)

pinyon (*P. edulis*)

scots (*P. sylvestris*)

stone (*P. pinea*)

Pistachio (*Pistacia*)

Plagioclimax (or disclimax)

Plantago (plantain, weed)

Pleistocene
Plough
Pluvial
 palaeoprecipitation
Podocarpus (southern yew)
Podzol; podsolization
Pollen
 analysis (or palynology)
 arboreal (AP)
 diagrams
 influx
 non-arboreal (NAP)
 zonation
Pollution, *see also* Acidification
 atmospheric
 eutrophication
Polynesian colonization of *see also* Maori
 Pacific
Post-glacial epoch *see* Holocene
Potato (*Solanum*), *see also* Sweet potato
Precipitates
Prehistory
Protocratic (or pre-temperate) substage
Proxy evidence

Rabbit (*Oryctolagus*)
Rackham, O.
Radiocarbon (^{14}C) dating
 AMS dating
 calibration of
Radiometric dating, *see also* Caesium-137; Lead(Pb), lead-210 dating
Ragweed (*Ambrosia*)
Reaves
Refuge areas
Reindeer (*Rangifer*)
Rice (*Oryza*)
Ritchie, Jim
Roman period
Ruddiman, Bill
 Early anthropogenic hypothesis
Rumex (dock)
Russia, *see also* Lakes, Baikal; Wrangel Island, Russian Arctic

Sahara desert
 early Holocene in Sahara
Sauer, C.O.
Savanna
Sclerophyll (hard-leaved)
Scotland
 Galloway lakes
 Late-Glacial period in
Sea level changes, *see also* Eustatic; Isostatic
Sedentism
Sediment,,
 flux
Shackleton, N.J.
Sheep (*Ovis*)
Sherratt, A.G.
Shifting (or swidden)
 cultivation
 Neolithic
 tropical
Shifting cultivation
Siberia
Simmons, I.G.
Site catchments
Slope deposit *see* Colluvium
Smith, A.G.
Sodium (Na)
Soils
 conservation
 erosion
 formation
 palaeosols (fossil or buried soils)
 salinization
Sorghum
South America, *see also* tropical rainforest
 Amazonia
 vegetation and climatic history of Andes
Southeast Asia
 sea-level history
Southern yew (*Podocarpus*)
Southwest Asia
Speleothems

Sphagnum moss
Spruce (*Picea*)
 decline in North America
Stadial
Steppe
Storrega slide *see* Tsunami
Stratigraphy, *see also* Peat
Street-Perrott, F.A.
Sub-Atlantic period (2600 Cal. yr BP to the present day)
Sub-Boreal period (5700-2600 Cal. yr BP)
Succession *see* Ecology
Suess, H.
Sulphur (S)
Sweden
 Lake Havgårdssjön and soil erosion
 pH history of Lake Oresjön
Sweet potato (*Ipomoea batatas*)
Swidden *see* Shifting, cultivation
Synchronous

Taphonomy
Taro (*Xanthosoma*)
Telocratic (or post-temperate) substage
Temperatures
 palaeotemperatures
 sea surface
 'thermal optimum' (hypsithermal)
Tephrochronology
Terminal Pleistocene *see* Late-Glacial (or terminal Pleistocene) stage (16,000–11,700 Cal. yr BP)
Testate amoebae
Thermocline
Thermoluminescence (TL) dating *see* Luminescence
Thermophilous (warmth-loving)
Thompson, L.G.
Till
Time-transgressive
Tree line
Tree rings *see* Dendrochronology
Troels-Smith, J.
Tropical rainforest

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Tsunami
 Storrega slide
Tundra
Turkey
 coastal evolution at Ephesus and Troy

Uniformitarianism *see also* Analogues
United States of America *see also* California
 climatic and vegetation history of Southwest
 Dendrochronology
 evolution of Great Lakes
 origins and development of North American prairie
 Palaeohydrology
 soil erosion and land-use,
 Frains Lake
Uranium-Thorium (U-Th) dating
Urination
 by packrats
 on tuber plants
Urtica (nettle)

van Andel, T.H.
van Zeist, W.
Varves (or laminations)
Vavilov, N.
Vera hypothesis
Viking, *see also* Greenland
Vita-Finzi, C.
Volcanism, *see also* Tephrochronology
Vostok ice core, Antarctica

Walker, D.
Watts, B. (AU: Not found)
Webb, T.
Weedy plants
Wetlands
Wheat (*Triticum*)
 Einkorn (*T. boeoticum* - wild, *T. monococcum* - domestic)
 Emmer (*T. dicoccoides* - wild, *T. dicoccum* - domestic)
Wilderness
Wind-blown, *see also* Aeolian

Windermere interstadial (14,600–12,900 Cal. yr BP), *see also* Lakes, Windermere

Bölling-Alleröd

Wisconsinan (Devensian) glaciation (ca.70-11,700 Cal. yr BP)

Wittfogel, K.A. *see also* Hydraulic civilization

Wolf (*Canis lupus*), *see also* Dog

Woodland *see also* Boreal, forest

ancient

Boreal forest

Deforestation

deciduous

evergreen

management

secondary

Wormwood/mugwort (*Artemisia*)

Wrangel Island, Russian Arctic

mammoths

Wright, H.

Xerophytic (drought-tolerant)

plants

Yams (*Diocorea*)

Younger Dryas stadial *see also* Dryas

Loch Lomond stadial